The

# Geophysical Journal

of the

Royal Astronomical Society

*incorporating*

*the Geophysical Supplement to the Monthly Notices of the R. A. S.*

**Volume 3 No. 1 March 1960**

*Published and Sold by the*

Royal Astronomical Society

Burlington House, London, W.1, England

*Price £1; in U.S.A. $3 Annual Subscription £3; in U.S.A. $9*

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*Editors*

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# The Compaction of Sediments and its Bearing on some Geophysical Problems

**D. S. Parasnis**

(Received 1959 April 20)

*Summary*

Experiments were undertaken to verify, under high pressures, the linear relation postulated by Terzaghi between the equilibrium void-ratio ($\epsilon$) of a sediment and the logarithm of the applied pressure ($p$). Samples of Gault, Oxford Clay, Keuper Marl, Globigerina Ooze and Terrigenous Mud were subjected to maximum pressures between 250 and 840 kg/cm$^2$. Certain departures were found from Terzaghi's relation. However, they do not appear to be such that large errors would arise in assuming the validity of Terzaghi's relation up to, say, 1 000 kg/cm$^2$. The void-ratio at unit pressure ($\epsilon_1$) was found to be proportional to the compression index (the coefficient of the logarithmic term in Terzaghi's relation). This suggests that if sediments reach a state of no voids, the pressure at which it is reached is the same for all of them. From the data obtained in the present experiments a pressure of about 2 500 kg/cm$^2$ is estimated for this state. Casagrande's suggestion that if an undisturbed pre-stressed sample of a sediment is re-subjected to pressure, the slope of the $\epsilon$–$p$ curve would show a discontinuity at the pre-stress value was verified under high pressure. An estimate of the overburden eroded from the top of the Athleta zone of the Oxford clay is made from such experiments.

Depth-density curves for the sediments concerned in the above experiments are presented. These calculations show that where compaction following Terzaghi's relation is the only process involved, the state of no voids is reached at about 10 000 m depth of burial.

Theoretical calculations of the heat conductivity of a mass of porous sediments are given in the paper. It is shown that the conductivity of the first 10 m of Globigerina Ooze, a typical deep sea sediment, should be expected to be reduced to two-thirds that of the constituent grains, owing to the presence of the water-filled voids.

The mechanism of compaction is discussed. An expression similar to Terzaghi's relation has been derived from certain considerations

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The mechanism of compaction is discussed. An expression similar to Terzaghi's relation has been derived from certain considerations

involving the energy needed to break the bonds between grains. The approach suggests that it should be possible to calculate the value of the compression index theoretically for a clay of known lattice structure.

## 1. Introduction

There are several geophysical problems where a knowledge of the manner in which sediments compact under high pressure would be desirable. For instance, the change of density with depth in sediments is bound to affect gravity anomalies over them. The velocities of seismic waves in sediments depend on the density while the heat flow depends on the porosity. In a qualitative way at least, the degree of compaction must be an important factor in such phenomena as mud flows, slumping and migration of oil. Again, if a certain formation has different thicknesses at different places in a given area, it does not necessarily follow that the rate of deposition varied in space. The facts can often be explained by "differential compaction". Jones (1942) has indicated that the present thickness of Edale shales in North Derbyshire is only 40 per cent of their original thickness. It is also a well known fact that if there are wedges of sand or sandstones in a mass of highly compressible sediments, differential compaction would give a dip of the series away from these discontinuities, thus affecting geological structure. We often encounter muddy sediments, which were once buried under a cover of overlying strata but are now very near the surface (e.g. Keuper Marl, Liassic Clays, etc.). If the law of compaction be ascertained for them it should be possible to estimate the thickness of the eroded rocks when the present degree of compaction is known.

Previous work on this subject has been chiefly carried out by soil physicists from the point of view of the stability of foundations on which large scale structures are built. The pressures involved in their studies have been moderate—up to about 100 kg/cm$^2$ or, in terms of the thickness of sedimentary strata of an average density 2·2 g/cm$^3$, about 500 m. The first systematic study of the porosity of rocks appears to be due to Sorby (1908), when he discussed the porosities of various limestones, shales and sandstones and derived a relation between the porosities of shales and their depth of burial. In 1920 Terzaghi in Vienna began a series of tests with his "Oedometer" for the laboratory determination of the consolidation constants of clays and in 1923 discussed mathematically the flow of water through a mass of clay. Ortenblad (1930) elaborated the theory. Athy (1930) and Hedberg (1936) derived depth–porosity relations from bore hole data. Mehl (1920), Teas (1923), Nevin (1929) and Trask (1931) have treated the differential compaction of sediments laid down on a basement of old erosional features. In 1944, Skempton confirmed the following relation (originally postulated by Terzaghi) for low pressures between $\epsilon$, the equilibrium void-ratio, and $p$, the pressure:

$$\epsilon = \epsilon_1 - c \log p/p_1 \qquad (1)$$

where $\epsilon_1$ and $c$ are constants for a given clay and $p_1$ is unit pressure. The void-ratio is the ratio of the pore space to the volume occupied by the mineral grains in a given sample.

The main object of the work described in the following sections was to extend the study of compaction to cover the high pressures corresponding to thicknesses of sediment of 4 000–5 000 m.

Freshly laid sediments contain a large amount of water but as new detrital load is added the pressure increases and causes the expulsion of water, bringing the individual grains closer together. This process of compaction is certainly an important factor in the transformation of soft sediments into hard rocks but it is, as will be seen later, not sufficient by itself. Chemical processes and metamorphism probably play a vastly more important role.

## 2. The apparatus and the technique of experiments

2.1. In the compaction of sediments on the floor of the ocean or in a deltaic deposit the horizontal linear dimensions are generally large compared to the vertical ones and with increasing pressure the water has presumably free access to other parts. The strains are only in the vertical direction, there being no lateral movement. In the laboratory these conditions were simulated by compressing a sediment confined in a steel cylinder. The specimen was placed between two porous disks so that the water freely drained out. The successive increments of pressure in nature are infinitesimal which is nearly impossible to achieve in the laboratory. In the author's experiments the pressures were raised by discrete amounts of 50–80 kg/cm$^2$, sometimes 150 kg/cm$^2$. Work on low pressure (Taylor 1942) suggests that $\epsilon_1$ and $c$ in equation (1) are essentially independent of the rate of pressure increment.

2.2. The apparatus consisted of a hydraulic ship jack with a robust steel frame round it. The frame was built of two mild steel plates (30 cm × 15 cm × 5 cm) held apart by four pillars. The deflection of the top plate was of the order of a few thousandths of a centimetre at a pressure of 4000 kg/cm$^2$ on the sample. As measurements of consolidation were not made with respect to any part of the plate, this deflection is of no consequence and in fact was helpful in partially retaining the pressure by a "spring effect". The pressure in the jack was multiplied about 8 times in the cylinder in which the sample was compressed and was applied manually with the jack's handle.

The cylinder and piston assembly for the compression of the sediment comprised: (a) a bottom-plate with a 1·3 cm high central boss on the upper surface, (b) a steel cylinder with an external diameter of 7·6 cm, an internal one of 3·49 cm, and a length of 12·1 cm, and (c) a 10 cm long piston fitting within a few thousandths of a centimetre in the cylinder. The bottom plate and the piston had two holes each (one vertical and one lateral) for the drainage of water. The cylinder had two diametrically opposite projections screwed into it on its outside. Their purpose will be made clear later.

The pressure on the sediment would drop to zero when the piston loses contact with the top plate of the frame as was bound to happen under progressive consolidation and the leakage in the pressure system. To counteract this loss of pressure an auxiliary pump was constructed which automatically pumped the necessary volume of oil into the jack when the pressure dropped from the pre-set value. The pump was driven by a small induction motor which was coupled to a simple servomechanism consisting of the Bourdon gauge for measuring the pressure, an adjustable electric contact and a relay. The motor started as soon as the pressure fell and stopped when it reached the pre-set value. The system kept the pressure constant to within 2 per cent over the entire range covered by the following experiments.

2.3. *The porous stones.* The sediment on which measurements were to be made was actually compressed between two porous disks of grinding stone (supplied by Messrs. Universal Grinding Wheel Co. Ltd). If the sediment were compressed directly the central holes in the piston and the bottom plate would be blocked by it. The disks were saturated with water and the bottom one was kept connected to a small water reservoir to prevent desiccation of the sediment through evaporation. No trouble which could be attributed to the blocking of the pores of these stones was experienced. Before this material was adopted, disks of sintered brass were tried but their pores were easily blocked up and could not be freed either mechanically or chemically.

Two grades of grinding stones were used. One was the standard type with an average pore size of 80 $\mu$m and the other, a specially manufactured one, with an average size of 14 $\mu$m. The former was found to be strong enough to withstand 1 000 kg/cm$^2$; the latter was pre-stressed to withstand 3 000 kg/cm$^2$. The structure of the standard grade is very open and it shows a permanent set in its virgin compression. Afterwards the compression is reversible.

2.4. *The dial gauge.* The consolidation of the sediment was read on a dial gauge attached to a vertical rod which was soldered to a ring. The ring could be adjusted anywhere on the piston while the prod of the gauge touched the bottom plate. A displacement of 0·0005 cm could be easily read on the gauge.

2.5. *The prevention of rusting.* In the trial experiments considerable trouble was experienced due to the rusting of the piston and the inside surface of the cylinder. It was important to prevent rusting because a large part of the applied stress goes in overcoming the friction due to it. A thin coating of an oil dag (colloidal graphite with oil base) applied by a brush along the entire inner surface of the compression cylinder was found completely effective in preventing rusting.

## 3. The individual compression curve

3.1. When an additional stress is applied to a clay already in equilibrium at a certain pressure, water is expelled and the clay continues to be compressed until it comes into equilibrium once more, with a new void-ratio at the new pressure. The curve obtained by plotting the dial readings against the time after the application of a new pressure will be called the "Individual Compression Curve". Such a curve is shown in Figure 1. From this the clay is seen to have come into equilibrium after about 24 hours, but if it is left under the pressure for a longer time it will continue to be compressed for several days at least. A sample of Northamptonshire "Iron clay" (Jurassic) 4 cm thick was left under a pressure of about 60 kg/cm$^2$ for nearly a week. The equilibrium was found to have been almost completely reached in the first 24 hours and only 1 per cent of the total compression took place in the next few days. Taylor's (1942) and Skempton's (1944) experiences concerning the time required for equilibrium are nearly the same.

Under very high pressures the individual compression curves of oozes appear to be discontinuous. This may depend upon the crushing of the grains or clusters of grains in the sample. At low to moderate pressures, the curves do not differ markedly from those of clayey sediments.

3.2. The flow of water through a porous packing is governed by Darcy's law which states that

$$v = \frac{k}{\eta}\frac{\partial p}{\partial z}, \tag{2}$$

where $v$ is the quantity of water per unit area per unit time, $k$ is a constant called the permeability, $\eta$ the viscosity and $\partial p/\partial z$ the pressure gradient across the packing.

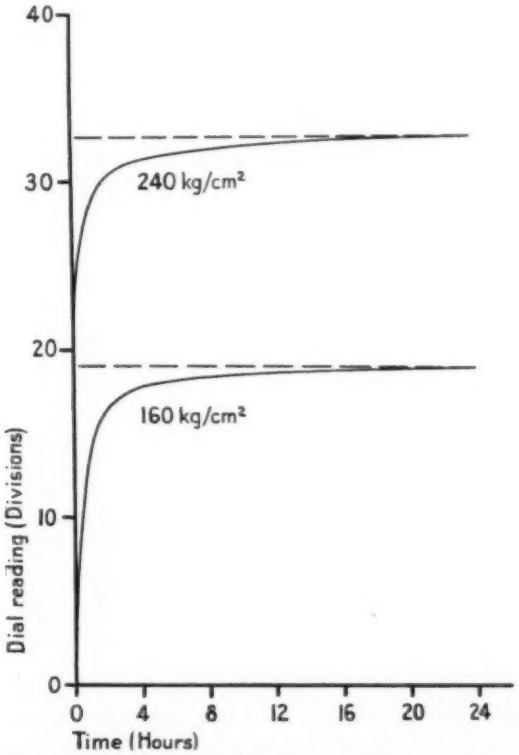


FIG. 1.—Individual compression curves at two successive pressures.

From equation (2), Terzaghi (1923) has derived the following relation for a packing resting on an impermeable, flat surface:

$$\frac{\partial u}{\partial t} = c_v \frac{\partial^2 u}{\partial z^2}, \tag{3}$$

where $u$ is the unbalanced hydrostatic pressure, $t$ is the time and $z$ the distance from the impermeable surface. The flow is supposed to take place along parallel lines perpendicular to the impermeable surface. The constant $c_v$ (cm²/s) is known as the coefficient of consolidation and is the analogue of "thermal diffusivity" in the theory of heat conduction. Terzaghi (1923) has treated the complete solution of (3) involving Fourier's series. In order to illustrate some features of the compaction of sediments in the present experiments, it will be sufficient to consider a

simple solution with the following boundary conditions:

(1) At $t = 0$, the excess hydrostatic pressure at $z = 0$ is equal to $\Delta p$, the pressure increment.

(2) At any time $t$, the excess hydrostatic pressure is zero at the two boundary surfaces $z = H$ and $z = -H$ where $2H$ is the thickness of the sample and the central plane $z = 0$ is the "impermeable" surface.

(3) At $z = 0$, $du/dz = 0$ for all $t$.

(4) When $t$ is very large, the excess hydrostatic pressure is zero for all $z$.

A solution found by separating the variables is

$$u = \Delta p\left\{\exp\left[-\frac{c_v}{H^2}\left(\frac{\pi}{2}\right)^2 t\right]\right\} \sin\frac{\pi}{2}\left(\frac{z}{H} + 1\right). \tag{4}$$

Now, if $v$ is the volume of water flowing across unit area in unit time $vdt$ is the consolidation of the clay in time $dt$. Therefore, in view of (2), the total compression in time $t$ is

$$\begin{aligned} S_t &= \frac{k}{\eta}\int_0^t \frac{\partial u}{\partial z}dt \\ &= S_\infty[1 - \exp(-p^2 t)], \end{aligned} \tag{5}$$

where

$$p^2 = \frac{c_v}{H^2}\left(\frac{\pi}{2}\right)^2$$

and $S_\infty$ is the value of $S_t$ when $t = \infty$. Equation (5) is the equation of the individual compression curve (Figure 1).

It is evident that $\mu = S_t/S_\infty$ is the same for all conditions for which $c_v t/H^2$ is the same. $c_v t/H^2$, which is dimensionless, has been called the "Time-factor" by Terzaghi.

3.3. An interesting consequence follows from the fact that $c_v t/H^2$ is dimensionless. Consider two thicknesses of sediments, for instance, $H_1 = 1$ cm and $H_2 = 3000$ m. Then if $c_v$ is the same, we have for the same degree of consolidation,

$$\frac{t_1}{H_1^2} = \frac{t_2}{H_2^2}. \tag{6}$$

If $t_1 = 24$ hrs, $t_2 \approx 200$ million years, which means that the consolidation of a 1 cm thick clay in 24 hours would be equivalent to that of 3000 m of sediment for about 200 million years. This is only an "order of magnitude estimate" and can be looked upon as providing justification for the interval allowed for compression in the present experiments, if they are to have a geological application. There are, of course, various objections to the argument. For instance, the sediments may be chemically affected during geological time or $c_v$ for laboratory samples may be very much larger than in nature. In the limiting case when $c_v$ for the latter is nearly zero, the result will be true for a very thin stratum only. But $c_v = 0$ means impervious sediments, a possibility which lies outside the scope of the present investigations. These factors may reduce $t_2$ possibly by a factor of 1000 or more but

nevertheless it would seem that the laboratory conditions represent a compaction of at least tens of thousands of years.

## 4. Variation of void-ratio with pressure

4.1. *The evaluation of void-ratio.* When a series of compression measurements on a sediment was completed, the sample was extracted, carefully cleaned and weighed immediately. It was then dried at 110°C to a constant weight in an oven and weighed after cooling in a desiccator. The grain density being known from a previous determination, the final void-ratio $\epsilon_f$ can be found from

$$\epsilon_f = \left(\frac{w_f}{w_d} - 1\right)\gamma_s, \tag{7}$$

where $w_f$ and $w_d$ are respectively the wet and the dry weights of the sample after extraction.

If $H$ is the thickness of the sample at any stage its void-ratio is given by

$$\epsilon = \frac{\pi r^2 H - V}{V},$$

where $r$ is the internal radius of the confining cylinder and $V$ the volume occupied by the grains. If we write $V = \pi r^2 H_0$,

$$\epsilon = \frac{H - H_0}{H_0}. \tag{8}$$

But

$$V = \frac{w_d}{\gamma_s}$$

so that $H_0$ is known.

If $H_f$ is the final thickness of the sample, then clearly

$$H_f = (1 + \epsilon_f)H_0. \tag{9}$$

The total compression $c$ due to a change of pressure can be found from the respective individual compression curves, and the values of $c$ can be successively added giving

$$H = H_f + \Sigma c, \tag{10}$$

$\epsilon$ is then calculated from (8).

4.2. *The corrections.* The estimation of the thickness of the clay sample is affected by the tilting of the piston and the dial gauge, oozing of some clay past the porous stones, the distension of the confining cylinder and the compression of the porous stones and the piston. It is estimated that for most observations in the present experiments the void-ratio is uncertain by, at the most, $\pm 5$ per cent.

The nominal gauge-pressure must be corrected for the friction between the inner surface of the cylinder and the sediment, since part of the applied pressure goes into overcoming this friction. Its value will depend on the length of the sample and the applied vertical pressure. Clearly, the force required to slide the cylinder over the sample is its measure. It will be recalled that the cylinder had two small

diametrically opposite projections screwed into it on the outside. One end of a simple lever consisting of two beams rigidly joined by two bolts rested against these projections while weights were added at the other end until the cylinder just slid. If $L$ is the lever ratio and $f$ the optimum weight, the thrust on the cylinder is $fL$ and the correction to the gauge reading is $fL/8{\cdot}3A$ where $A$ is the area of the sample and 8·3, the pressure-multiplication factor. The friction cannot be found at the same time as the main experiment because the sediment would be disturbed. It was measured in a separate experiment on a sample under identical conditions. Figure 2 shows some typical observations. Since the friction experiment could not be carried out safely for very high pressures the corresponding correction was found by the extrapolation of straight lines such as those in Figure 2.

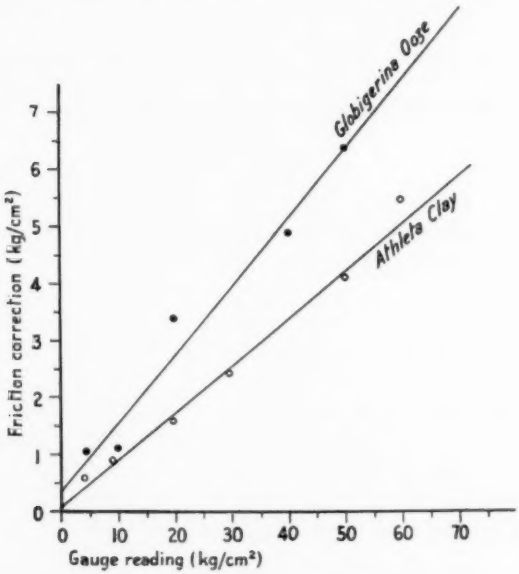


FIG. 2.—Friction correction.

4.3. *Observations on slurried sediments.* In this class of experiments, a sample was thoroughly mixed with water to form a homogeneous slurry and some of it was dropped (and worked in place by a spatula) into the cylinder. At the same time a portion out of the same stock was transferred to a beaker for determining the water content. The sample may be straight from the earth, in which case the slurrying is called primary, or it may be the "claystone" produced artificially under high pressure, in which case the slurrying is called secondary. This compressed sample can be re-slurried and thus a tertiary slurrying is effected, and so on. In the primary slurry, all particles greater than about $\frac{1}{2}$ mm were generally removed during preparation.

The sample extracted from the cylinder after a compression series was a tough, compact structure which has already been referred to as "claystone". In appearance it was not unlike mudstone or siltstone, though not so consolidated. When dry it was brittle and a little harder than the finger nail, the hardness varying with the material.

The sediments on which observations were made fall into two classes:

1. Old consolidated sediments which included
   a. Gault
   b. Oxford Clay
   c. Keuper Marl
2. Recent Deep Sea sediments which included
   a. Globigerina Ooze from various parts of the Atlantic Ocean
   b. Terrigenous Mud

The first class is clayey, while the second is silty. The latter have little or no "bonding" between the particles and exhibit only slight plasticity.

For each sample investigated the following relevant physical characteristics were determined: average grain density, liquid limit, calcium carbonate content and particle size. They will be found in Table 1.

The liquid limit of a clay is defined as that water content at which a 1 cm gap in the clay just closes when 25 blows are imparted to it in the standard Casagrande apparatus. This is an arbitrary measure but serves as a useful index property. Physically, it represents a condition in which the clay is on the verge of plasticity and liquidity.

The particle size distribution was determined by the method of drawing aliquots at fixed known intervals from a weak, slightly alkaline suspension of the clay in question. The method used was a slight modification of that described by Dallavalle (1948).

4.3.1. *Gault.* The Gault sample was obtained by boring in the southwest field of the Cambridge Observatory. It was encountered at a depth of about 5 ft after a top soil layer followed by some arenaceous material. The boring technique was not capable of giving an undisturbed sample. The particular Gault obtained is a very stiff, dark green or blue clay with occasional small pebbles.

*Natural water content:* 26 per cent corresponding to a void-ratio of 0·690.

*Results of measurements:* Two series of observations were taken on primary slurries. The length of the sample in series 2 was about double that in the first. The results are plotted in Figure 3. The slight systematic difference between the two series may be due to different initial water contents (*cf.* Skempton 1944). The straight lines in Figure 3 have the following equations:

$$\textit{Series 1} \quad \epsilon = (1{\cdot}10 \pm 0{\cdot}012) - (0{\cdot}277 \pm 0{\cdot}005) \log_{10} p/p_1,$$

$$\textit{Series 2} \quad \epsilon = (1{\cdot}00 \pm 0{\cdot}028) - (0{\cdot}241 \pm 0{\cdot}006) \log_{10} p/p_1,$$

where $p$ is in kg/cm$^2$ and $p_1$ is unit pressure.

Two further curves (secondary and tertiary slurrying) on the sample derived from series 2 are also plotted in Figure 3. They are not straight lines obeying Terzaghi's relation. For pressures greater than about 30 kg/cm$^2$ the primary, secondary and tertiary curves give successively lower void-ratios.

4.3.2. *Oxford Clay* (*Athleta zone*). Undisturbed samples from various depths in the pits at Stewartby (near Bedford) were supplied by the London Brick Company. The Athleta is a well stratified, almost shaly, very stiff, dark, abundantly

fossiliferous clay giving typical conchoidal fractures. The void-ratio of the surface sample was 0·750.

*Results of measurements* (Figure 4). A straight line fitted by considering all the points on the primary curve gives

$$\epsilon = (1{\cdot}549 \pm 0{\cdot}073) - (0{\cdot}442 \pm 0{\cdot}028) \log_{10} p/p_1.$$

However there is a slight, but significant, curvature and neglecting the two points at the lowest pressures,

$$\epsilon = (1{\cdot}233 \pm 0{\cdot}120) - (0{\cdot}333 \pm 0{\cdot}043) \log_{10} p/p_1.$$

The secondary curve does not cover the same pressure range as the primary, but the tendency towards lower void-ratios at the same pressure is apparent.

4.3.3. *Keuper Marl.* The sample was obtained from the Aust cliff near the mouth of the Severn. It is a fairly compact, friable rock with a deep red or chocolate appearance. The particular sample was from the uppermost Rhaetic beds just below the "Tea-green Marls". It is difficult to slurry Keuper Marl. The natural void-ratio of the sample was 0·22.

*Results of measurements* (Figure 4):

$$\epsilon = (0{\cdot}540 \pm 0{\cdot}0004) - (0{\cdot}115 \pm 0{\cdot}0019) \log_{10} p/p_1.$$

4.3.4. *Globigerina Ooze.* Three samples were available, two of which were supplied by the British Museum of Natural History and one by the Department of Geology, Columbia University. Naturally, very limited amounts were available.

Sample No. 1

This came from station No. 338 (Lat. 21°15′ S, Long. 14°2′ W) of the H.M.S. *Challenger* expedition. (Depth 3 640 m.) In the reports of the expedition the following analysis is given:

| | |
|---|---|
| Calcium carbonate | 92·54 per cent (85 per cent Pelagic foraminifera) |
| Minerals | 1·00 per cent |
| Siliceous remains | 1·00 per cent |
| Fine washings | 5·64 per cent. |

The minerals are mentioned to have a median diameter of 0·06 mm. They consist of hornblende, felspars, magnetite, pumice, manganese etc. The general appearance of the ooze is "white with rosy tinge, granular, homogeneous, chalky". Under the microscope the ooze was found to consist of particles of various shapes and sizes with spherical particles as an exception.

*Results of measurements* (Figure 5):

$$\epsilon = (2{\cdot}591 \pm 0{\cdot}036) - (0{\cdot}734 \pm 0{\cdot}015) \log_{10} p/p_1.$$

Sample No. 2

This came from a depth of 1 257 m at a spot 50°13′ N, 11°23′ W. It had a lighter appearance than Sample 1.

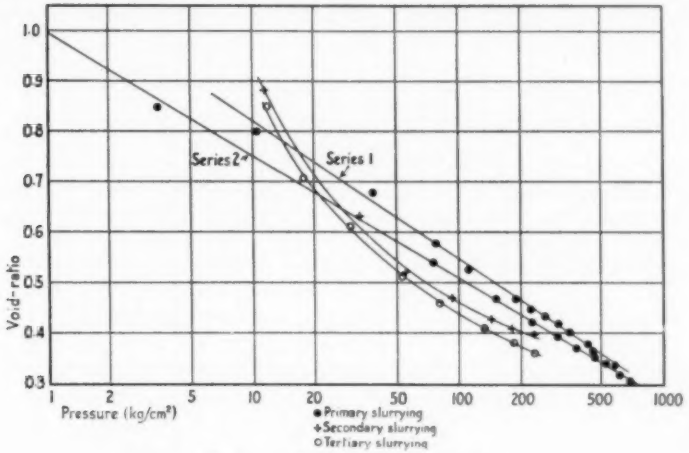


FIG. 3.—Compaction of Gault.

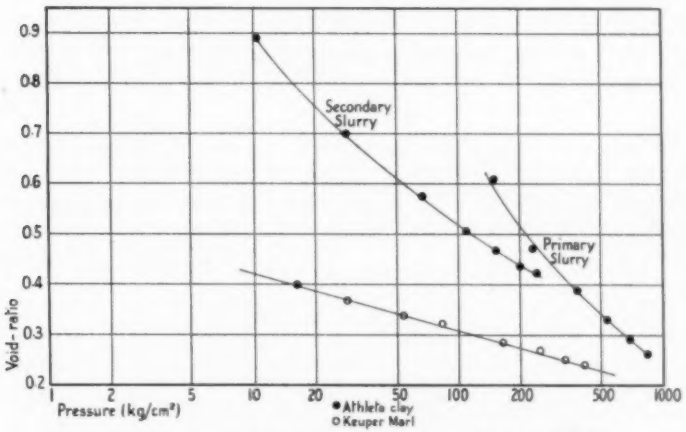


FIG. 4.—Compaction of Athleta clay and Keuper Marl.

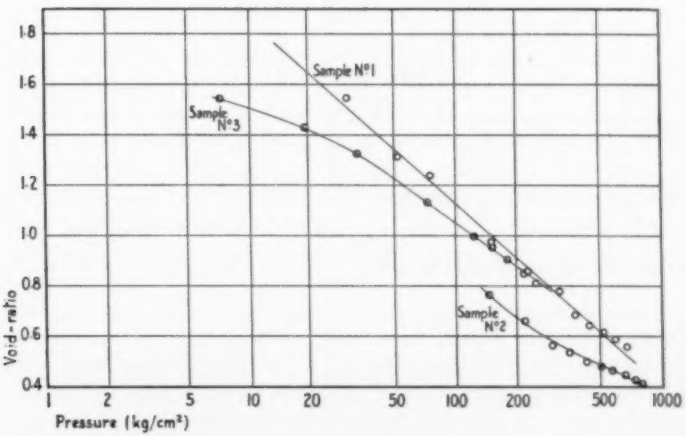


FIG. 5.—Compaction of Globigerina Ooze.

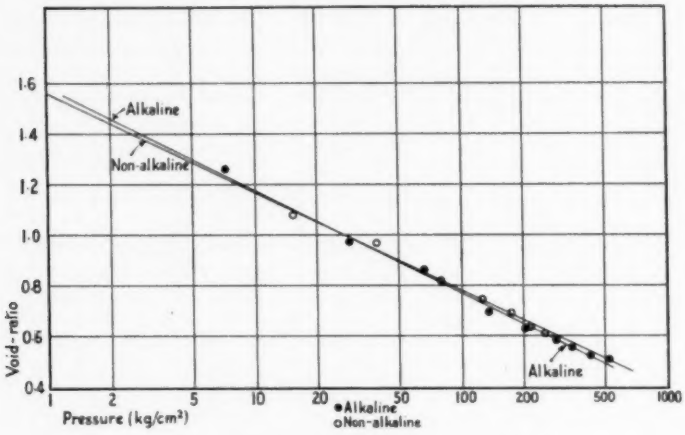


FIG. 6.—Compaction of Terrigenous Mud.

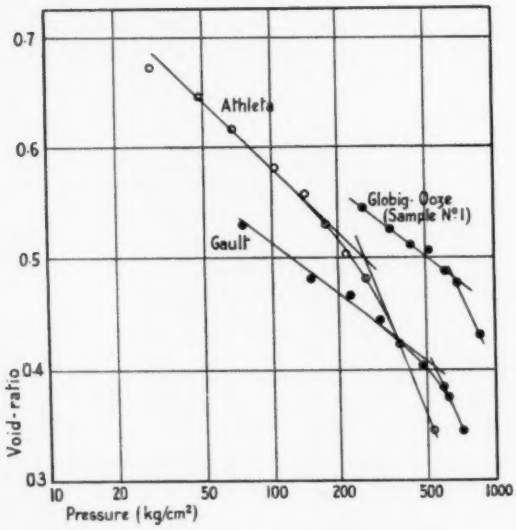


FIG. 7.—Compaction of pre-stressed samples.

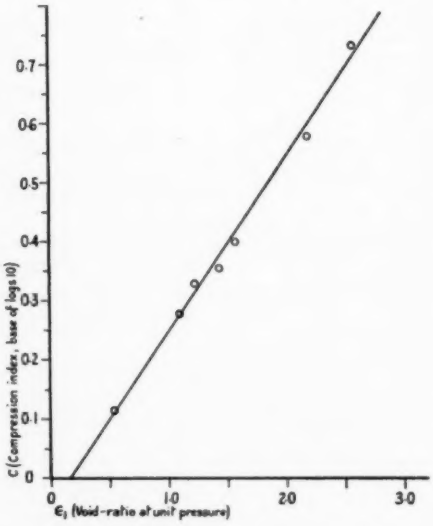


FIG. 8.—Relation between $C$ and $\epsilon_1$.

*Results of measurements* (Figure 5). The observed points do not fall on a single straight line. On neglecting the two points at the two lowest pressures, the equation of the best straight line is

$$\epsilon = (1{\cdot}440 \pm 0{\cdot}027) - (0{\cdot}356 \pm 0{\cdot}010) \log_{10} p/p_1.$$

Sample No. 3

This came from a depth of 3480 m, 0·8 m below the surface of the sediment at a spot 30°04′ N, 41°25′ W.

*Results of measurements* (Figure 5). The points for pressures greater than 30 kg/cm² fall on a straight line with the equation

$$\epsilon = (2{\cdot}203 \pm 0{\cdot}008) - (0{\cdot}579 \pm 0{\cdot}004) \log_{10} p/p_1.$$

The initial part of the $\epsilon - \log p$ curve is flatter than the above straight line, an observation which is contrary to those on Sample 2 as well as Gault and Oxford clay.

4.3.5. *Terrigenous Mud.* The sample came from a depth of 4700 m at a spot 45°26′ N, 9°20′ W, that is, from approximately the same part of the Atlantic Ocean as sample No. 2 of Globigerina Ooze. It was described as "impure" and contained small glass fragments which were removed before the compaction experiments.

*Results of measurements* (Figure 6). Two series were taken, one with the mud mixed in highly alkaline water and the other in ordinary tap water. They agree within the experimental errors, a result suggesting that ionic exchanges do not play an important part in this case. The straight lines in Figure 6 can be represented by the following equations:

Alkaline

$$\epsilon = (1{\cdot}587 \pm 0{\cdot}017) - (0{\cdot}406 \pm 0{\cdot}008) \log_{10} p/p_1,$$

Non-alkaline

$$\epsilon = (1{\cdot}558 \pm 0{\cdot}028) - (0{\cdot}389 \pm 0{\cdot}014) \log_{10} p/p_1.$$

The difference between the corresponding parameters of the two equations is not significant.

4.4. *Observations on pre-stressed samples.* The compression of a stratum of clay is irreversible but under very high pressures, a part at least of the compression will be elastic due to the recovery of the grains themselves. If a specimen of clay is stressed to a certain pressure and the pressure is continuously released, the change in the void-ratio will be very small so that the $p$–$\epsilon$ curve of "recovery" will be almost parallel to the $p$-axis. Casagrande suggested that if such a sample is stressed again the rate of change of $\epsilon$ with pressure will suddenly assume a higher value when the point of pre-stressing is reached. Pre-stressed samples can be obtained from nature or can be made in the laboratory.

The results of experiments undertaken to verify Casagrande's suggestion under high pressure are shown in Figure 7. The Gault sample was prestressed in the laboratory to a pressure of 580 kg/cm². The Globigerina Ooze sample was the one extracted after the experiments represented by Figure 5 (Sample 1). The samples were resaturated without disturbing the structure and compressed again.

Although a sufficient number of points are not available in Figure 7 after 500 kg/cm$^2$ the change in slope of the $\epsilon - \log p$ curves is clearly seen to take place at the pre-stress value.

An undisturbed sample from natural strata will be pre-stressed to the pressure of the overburden whether it is now present or eroded. Obviously, we have here a tool for the determination of eroded thicknesses. The difficulty of course is of obtaining a completely undisturbed sample since some disturbance due to the coring instrument is always present. For the author's experiments, a good undisturbed sample of Athleta clay was fortunately supplied by the London Brick Company. A specimen was very carefully cut from a core 15 cm long and 7·5 cm in diameter so as to fit in the compression cylinder. From the result shown in Figure 7 it can be seen that a change of slope in the $\epsilon - \log p$ curve occurs at about 280 kg/cm$^2$. This pressure corresponds to a thickness of about 1 400 m of strata of an average density of 2·0 g/cm$^3$ (see further Section 6·2).

## 5. Interpretation of the results

5.1. *The validity of Terzaghi's relation.* The experiments summarized above show that Gault and Keuper Marl follow Terzaghi's relation up to about 700 kg/cm$^2$ and 400 kg/cm$^2$ respectively, but Athleta clay shows significant departures. The Globigerina Ooze samples seem to obey the relation (for sufficiently high pressures) to about 800 kg/cm$^2$ while Terrigenous Mud satisfies it from very low pressures to pressures of at least 500 kg/cm$^2$. Repeated slurrying seems to lead to lower void-ratios at the same pressure and also to a significant departure from Terzaghi's relation.

The observed departures from Terzaghi's relation are such that $c$ in equation (1) is smaller at higher pressures. The one exception observed here is Globigerina Ooze, Sample 3. If the variation of $c$ is continuous, the curvature of the $\epsilon - \log p$ curves can be reduced, if a constant empirical quantity $\psi$ be subtracted from the nominal applied pressure. In fact $\psi$ may be chosen by the method of least squares to give a minimum curvature. This consideration implies that

$$\epsilon = \epsilon_1 - c \log(p - \psi)/p_1 \qquad (11)$$

($\psi$ will be negative for the exceptional sample of Globigerina Ooze).

It may be mentioned that a systematic error in the reading of the pressure gauge will also give rise to equation (11). However no such error sufficient to explain the curvatures in Figures 3, 4 and 5 is suspected in the present experiments. The pressure gauges employed were tested by a Standard Gauge Tester and found to be accurate to within 1 per cent. The errors in the friction correction will be of either sign and therefore should not give rise to equation (11). The friction correction, in any case, is estimated to be correct to within $\pm 3$ kg/cm$^2$.

It is, however, conceivable that the curves in Figures 3, 4 and 5 not following Terzaghi's relation actually consist of two parts, each being a straight line. In this case, the pressure at which the discontinuity in slope occurs may be a characteristic of the crushing strength of the grains or some quantity of that nature.

Table 1 lists, among other data, the values of $\epsilon_1$ and $c$ for the samples tested, in those cases where the results could be fitted to equation (1).

5.2. *Correlation of c and calcium carbonate content* ($w$). If calcium carbonate is present in a sediment in appreciable amounts the crushing of the grains would be

expected to influence the compression index. The data in Table 1 yield the following correlation.

$$c = (0{\cdot}178 \pm 0{\cdot}088) + (4{\cdot}10 \pm 1{\cdot}45)\,10^{-3} w. \tag{12}$$

Statistical examination suggests that the correlation is significant. However more observations are needed before asserting it with confidence.

5.3. *Correlation of* $\epsilon_1$ *and* $c$. Skempton (1944) has found a linear relation between $\epsilon_1$ and $c$. An important consequence follows from this relation. For let $\epsilon_1 = \eta c$, then

$$\epsilon = c(\eta - \log p).$$

Hence $\epsilon = 0$ when $\log p = \eta$, a constant for all clays, which means that if clays reach a state of "no voids" (perfect compaction), the pressure at which it is reached is the same for all clays. But if $\epsilon_1 = \eta c + b$ then obviously,

$$\epsilon = 0 \text{ when } \log p = \eta + \frac{b}{c}$$

which means that this pressure is dependent on $c$ only. Figure 8 shows the graph of $\epsilon_1$ and $c$ for the author's experiments. The equation of the best straight line is

$$\epsilon_1 = (3{\cdot}39 \pm 0{\cdot}12)c + (0{\cdot}17 \pm 0{\cdot}05), \tag{13}$$

$b$ is significantly different from zero but only barely so on the given degrees of freedom. If we assume for simplicity that it is zero, then $\epsilon = 0$ for $p \approx 2\,500$ kg/cm$^2$.

**Table 1**

*Data on sediments tested under high pressure*

| Specimen | $\epsilon_1$ | $c$ | Particle size <20μm silt + clay | Particle size <2μm "clay" | $CaCO_3$ per cent by weight | Average grain density g/cm$^3$ | Liquid limit per cent | Max. test pressure kg/cm$^2$ |
|---|---|---|---|---|---|---|---|---|
| Gault | | | | | | | | |
| Sample (1) | 1·10 | 0·277 | 60·0 | 16·7 | 36·6 | 2·653 | 85·0 | 580 |
| Sample (2) | 1·00 | 0·241 | 60·0 | 16·7 | 36·6 | 2·653 | 85·0 | 680 |
| Athleta Clay | 1·23 | 0·33 | 65·0 | 20·0 | negligible | 2·450 | 72·7 | 840 |
| Keuper Marl | 0·54 | 0·115 | 64·0 | 26·8 | 28·35(d) | 2·712 | 23·6 | 420 |
| Globigerina Ooze | | | | | | | | |
| Sample (1) | 2·59 | 0·734 | (a) | (a) | 92·54(b) | 2·650 | (a)(c) | 670 |
| Sample (2) | 1·44 | 0·356 | (a) | (a) | 48·0 | 2·635 | (a)(c) | 780 |
| Sample (3) | 2·20 | 0·58 | 15·0 | negligible | 97·30 | 2·590 | (a)(c) | 250 |
| Terrigenous Mud | 1·57(e) | 0·398(e) | 50·0 | 0·3 | 56·0 | 2·670 | ca 50 | 520 |

Notes: (a) Sufficient amount of specimen was not available for this determination.
(b) Determined by the H.M.S. *Challenger* expedition.
(c) The liquid limit of a silty sediment like Globigerina Ooze is very low.
(d) This should be viewed with caution because Keuper Marl is known to contain occasionally an appreciable amount of dolomite.
(e) Mean of the curves for alkaline and non-alkaline media.

## 6. Compaction and some geophysical problems

6.1.1. *Depth–void-ratio and depth–density relations.* If $\gamma$ is the "bulk" density of a sediment at a depth $z$, then the pressure of the overburden is

$$p = \int (\gamma - 1)dz. \tag{14}$$

The factor $(\gamma - 1)$ occurs because when the pores of a sediment are freely intercommunicating, the head of the water of saturation gives a buoyancy to the grains. If the pores are freely intercommunicating to a depth of $z_h$ only, then the pressure at a depth $z$ $(> z_h)$ will be

$$p = \int_0^{z_h} (\gamma - 1)dz + \int_{z_h}^{z} \gamma dz. \tag{15}$$

The void-ratio at which hydrostatic uplift largely ceases appears to be about 0·5 from Terzaghi's experiments (1936).

Depth–void-ratio relations have been found empirically (Athy 1930, Hedberg 1936), but from equation (14) combined with equation (1) a general relation can be derived. The solution of the appropriate first order differential equation involved is

$$z(\gamma_s - 1)\exp(-\epsilon_1/C)/p_1 = (1 + \epsilon + C)\exp(-\epsilon/C) + K, \tag{16}$$

where $C$ = compression index (base $e$) = $c/2{\cdot}303$; $\gamma_s$ = grain density; $K$ = constant of integration.

Cook (1952) determines $K$ under the assumption that the void-ratio of a newly deposited clay ($z = 0$) is equal to $\epsilon_l$, the void-ratio at the liquid limit. Two alternative assumptions can be made, and justified in particular cases, for the value of $K$. (i) $K = 0$ which is equivalent to assuming an infinite void-ratio (very dispersed grains) for $z = 0$, and (ii) $K = -(1 + \epsilon_1 + C)\exp(-\epsilon_1/C)$ which is equivalent to assuming the effective pressure on the sediment to be unity (approximately equal to the atmospheric pressure) for $z = 0$. The $z - \epsilon$ curves obtained on the three different assumptions differ markedly from each other in the first few tens of metres only. For instance, for Cook's clay 3($\epsilon_1 = 0{\cdot}7$, $C = 0{\cdot}09$, liquid limit = 30), the calculations with $K = 0$ give a depth which is throughout 3·3 m larger than the corresponding value on Cook's assumption. The difference becomes proportionately less significant as the depth increases.

The results of a calculation on Globigerina Ooze (Sample 1, $\epsilon_1 = 2{\cdot}59$, $C = 0{\cdot}319$) assuming $K = 0$ are shown in Figure 9.

The problem of the depth–density relation is not essentially different since the "bulk" density $\gamma$ is equal to $(\gamma_s + \epsilon)/(1 + \epsilon)$.

Some help in quickly converting $\epsilon$ into $\gamma$ is afforded by the nomogram in Figure 10. The depth–density curves for Keuper Marl, Gault, Terrigenous Mud and Globigerina Ooze are shown in Figure 11.

The depth for $\epsilon = 0$ can be easily found from equation (16). This estimate will be based on the assumption that hydrostatic uplift of the grains is effective at all stages. If this be not true we must find the $z$-$\epsilon$ relation by using equation (15)

instead. Assuming tentatively that $z_h$ corresponds to $\epsilon = 0{\cdot}5$, its value can be found from equation (16). The second term of equation (15) gives

$$\frac{dp}{dz} = \gamma = \frac{dp}{d\epsilon}\frac{d\epsilon}{dz},$$

substituting for $dp/d\epsilon$ from Terzaghi's relation, rearranging the terms and integrating between $\epsilon = 0$ and $\epsilon = 0{\cdot}5$ we get:

$$A\int_{z_h}^{z_p} dz = \int_0^{0.5} \exp(-\epsilon/C)d\epsilon - (\gamma_s - 1)\int_0^{0.5} \frac{\exp(-\epsilon/C)}{\gamma_s + \epsilon}d\epsilon, \qquad (17)$$

where

$$A = C/p_1 \exp(-\epsilon_1/C).$$

$z_p$ denotes the depth at which $\epsilon = 0$, i.e. the state of "perfect compaction". To integrate the second term on the right-hand side of (17), $1/(\gamma_s + \epsilon)$ can be expanded in ascending powers of $\epsilon/\gamma_s$ which is less than unity in the range of integration. Retaining only the first power of $\epsilon/\gamma_s$, evaluating the definite integrals and simplifying the resulting expressions, we get:

$$A(z_p - z_h) = (C/\gamma_s)[1 + C - C/\gamma_s - \{1{\cdot}500 + C - (C + 0{\cdot}5)/\gamma_s\}\exp(-0{\cdot}5/C)]. \qquad (18)$$

Obviously, the value of $z_p$ is obtained by adding $z_h$ and $z_p - z_h$ as calculated from equations (16) and (18) respectively. Table 2 gives the results of such calculations on Keuper Marl, Gault, Terrigenous Mud and Globigerina Ooze (Sample 1). For comparison, the depth of perfect compaction $z'_p$, assuming buoyancy of the grains at all stages, is also included. For $\epsilon_1/C$ the value obtained from equation (13), namely 7·81, has been used.

**Table 2**

*Values of $z_p$ and $z_p'$*

| Sediment | $C$ | $z_h$ m | $z_p - z_h$ m | $z_p$ m | $z'$ m |
|---|---|---|---|---|---|
| Keuper Marl | 0·050 | 10 | 9550 | 9560 | 15400 |
| Gault | 0·113 | 260 | 9675 | 9935 | 16350 |
| Terrigenous Mud | 0·173 | 1290 | 9425 | 10715 | 17250 |
| Glob. Ooze | 0·319 | 5550 | 8125 | 13675 | 19350 |

It is interesting to note that the depth at which zero void-ratio is attained is about 40–60 per cent greater when hydrostatic uplift is effective at all stages than when it is effective up to $\epsilon = 0{\cdot}5$ only. In the latter case the depth will, of course, depend on the critical value of $\epsilon$ which may be different for different sediments.

6.1.2. *The error* $\Delta(z_p - z_h)$ in neglecting powers of $\epsilon/\gamma_s$ higher than the first in deriving (18) can be easily estimated. The integral involving the Taylor remainder gives,

$$A\Delta(z_p - z_h) = \int_0^{0.5} \exp(-\epsilon/C)(\epsilon^2/2\gamma_s^2)f''(\theta\epsilon/\gamma_s)d\epsilon, \qquad (19)$$

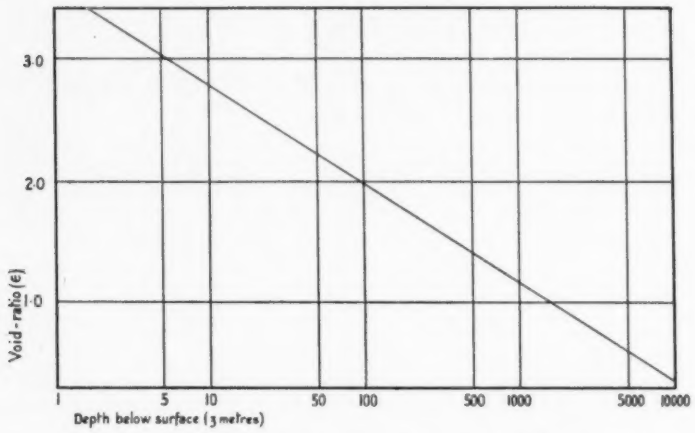


FIG. 9.—$e$-$z$ relation for Globigerina Ooze (sample No. 1).

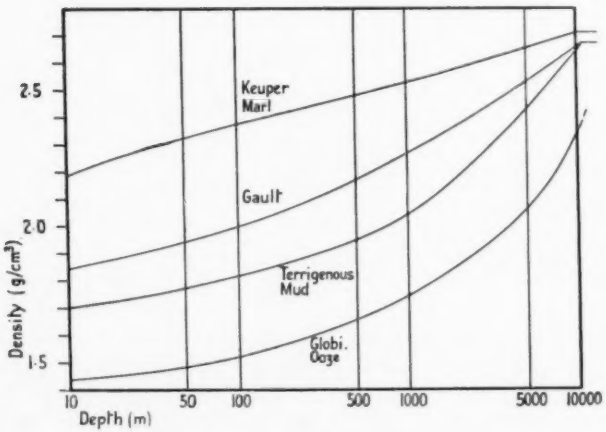


Fig. 11. Depth–density curves.

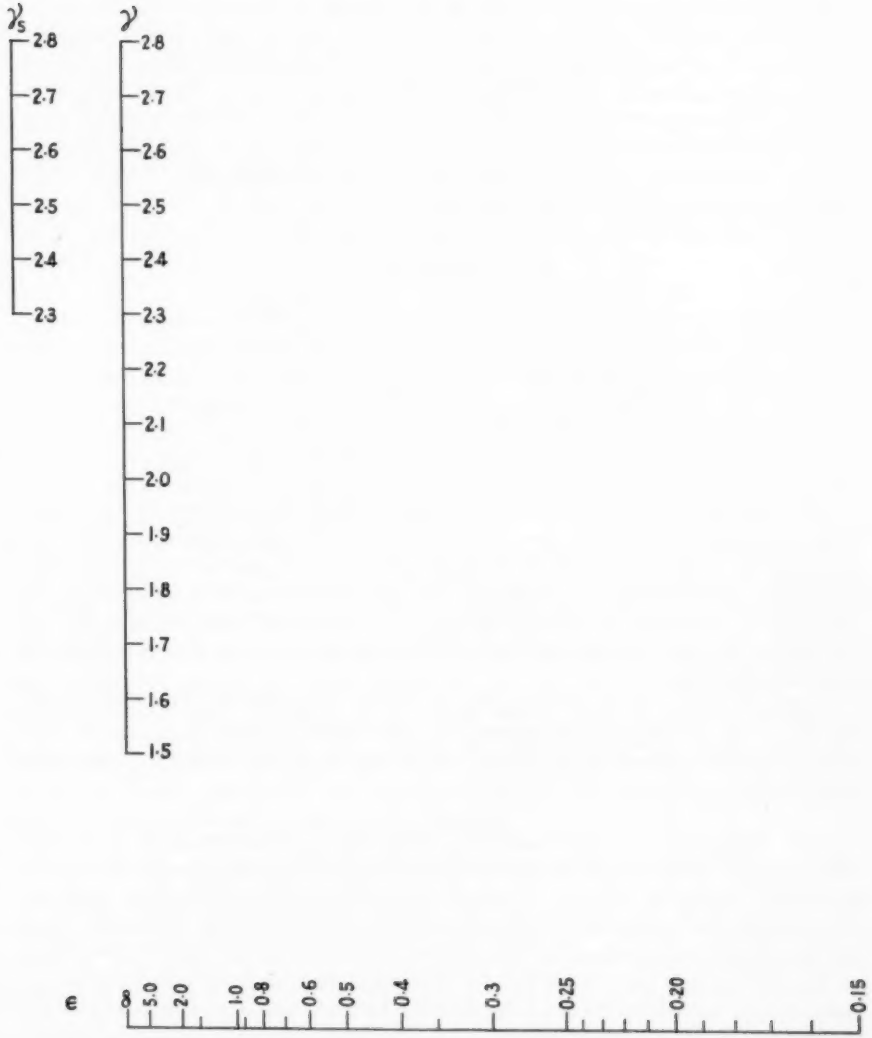


FIG. 10.—Nomogram for quick calculation of the density of a sediment ($\gamma$) from the void-ratio ($\epsilon$).
(A straight edge cutting the three axes gives the mutually consistent values of $\epsilon$, $\gamma_s$ and $\gamma$).

where

$$0 \leqslant \theta \leqslant 1 \quad \text{and} \quad f''(\epsilon/\gamma_s) = 2\gamma_s^{-2}(1+\epsilon/\gamma_s)^{-3}.$$

The maximum value of $f''(\theta\epsilon/\gamma_s)$ is $2\gamma_s^{-2}$. Hence,

$$\Delta(z_p - z_h) < A^{-1} \int_0^{0.5} \gamma_s^{-4}\epsilon^2 \exp(-\epsilon/C)d\epsilon$$

$$= A^{-1}C\gamma_s^{-4}[2C^2 - (0.25 + C + 2C^2)\exp(-0.5/C)]. \quad (20)$$

The values of $\Delta(z_p - z_h)$ are listed in Table 3.

**Table 3**

| Sediment | $[\Delta(z_p - z_h)]_{\max}$ m |
|---|---|
| Keuper Marl | 3 |
| Gault | 10 |
| Terrigenous Mud | 16 |
| Globigerina Ooze | 21 |

The depths for $\epsilon = 0$ in Table 2 are calculated on the assumption that such a state is reached for the sediments in question. However this assumption may not be always justified. For instance, a clay composed of *uniform spherical* grains has a minimum void-ratio of 0·35 so long as the spheres are not deformed. Then the lower limit of integration in (17) will be 0·35. Such cases are probably only of theoretical interest and therefore no further calculations will be given (see however section 7.1 below).

Depth–density curves such as those in Figure 11 should be useful in the estimation of the gravitational attraction of a clay deposit. Cook (1952) has already considered this problem in great detail for clays with certain assumed compression constants.

6.2.1. *Estimation of eroded overburden.* In section 4.4 it was pointed out that the discontinuity in the slope of the $\epsilon - \log p$ curve of an undisturbed clay sample provides a means of estimating eroded overburden. For the Athleta clay the slope changes at 280 kg/cm² (Figure 7). To find the thickness of eroded strata corresponding to this pressure we assume that the original sedimentation curve of the Athleta was close to that of the slurried laboratory sample (Figure 4). Accordingly, the void-ratio at 280 kg/cm² will be 0·45, and from equation (16) the thickness of the overlying sediments is found to be 1 150 m under the assumption inherent in equation (16) that the grains experience a hydrostatic uplift.

Geological data concerning the thickness eroded from the top of the Athleta clay are very meagre. Nevertheless, certain estimates can be made on the following lines. The particular laboratory sample came from the pits near Stewartby in the Lower Jurassic system. During the Upper Jurassic and the Lower Cretaceous the Midlands were uplifted owing to the Saxonian orogenic movements postulated by Stille. This re-elevation took place in the region where the Mercian Highlands of the Trias times stood and continued from nearly Pre-Portland to Gault times. We shall assume that all the strata approximately from the Corallian (Jurassic) to the

Gault (Cretaceous) have been eroded. That all of them except the Portlandian and the Purbeckian were deposited here seems quite likely (*cf.* Regional Handbook, Central England District). By analogy with the Hampshire district, the full development of the Cretaceous i.e. only the Wealden for our district, would have a thickness of about 760 m and that of the Corallian about 60 m. Between these two deposits occurs the Kimmeridgian which is nearly 370 m thick in the Hampshire district but only about 30 m in parts of Oxfordshire. It is difficult to arrive at a reliable figure applicable to the region around Stewartby. With due regard to the two extremes, it seems that the thickness of the strata eroded from the top of the Athleta clay is probably between about 790 and 1 130 m. These figures agree closely with the "compression estimate" above. Clearly, a number of physical and geological assumptions are involved in the two estimates and the close agreement can be fortuitous.

6.2.2. If an undisturbed sample of a clay is not available we can attack the problem of the eroded overburden in a slightly different way. According to Cooling & Skempton (1941) and Jones (1944) the curves for undisturbed and remoulded samples tend to coincide under high pressure. A comparison of Figure 10(b) with Figure 4 and Figure 10(c) with Figure 5 (Sample 1) will support this. Thus we shall be justified in using the "remoulded" instead of the "undisturbed curve" provided the pressures in question are sufficiently high. Then if the void-ratio is known, the pressure at which it is attained can be read off the $\epsilon - \log p$ curve. For instance, Keuper Marl (from the Aust Cliff) has a surface void-ratio of 0·22 which, from Figure 7, is seen to correspond to a pressure of about 500 kg/cm$^2$. Hydrostatic uplift is very probably not effective at this void-ratio. Hence the above pressure would correspond to that of a 2 500 m thick overburden with an average density of 2·0 g/cm$^3$ under which the Keuper Marl beds consolidated and which is now eroded. A reliable geological check on this estimate is not available. The nearest district where the thicknesses of the post-Rhaetic strata are known in detail is again the Hampshire basin. From the information given in the Regional Handbook a maximum thickness of about 2 200 m can be deduced for the post-Rhaetic in this district. To accept this figure as a check on the laboratory estimate would be tantamount to assuming the following:

(1) The development in the Hampshire basin extended northwards at least as far as the present Aust Cliff.
(3) The Saxonian orogeny was effective near the present Aust Cliff.
(3) The average bulk density of all the beds from the Recent to the Liassic was 2·0 g/cm$^3$.

6.3. *Heat conductivity of porous sediments.* It is proposed here to calculate the conductivity of a sediment with a given water content by taking a definite model for the aggregate. Four models naturally suggest themselves: (1) the grains in the sediment are long, thin plates oriented perpendicular to the lines of heat flow, (2) the grains are long, cylindrical bodies with the axis of the cylinder perpendicular to the same lines, (3) the grains are spheres, (4) the grains are long, thin plates with their lengths parallel to the lines of heat flow. The last model, does not, *a priori*, seem likely because the grains would naturally lie horizontally as the pressure of the overlying sediment increases. The effective conductivity of a sediment can be calculated by a self-consistent method which is an extension of that used by

McKenzie (1950) for the calculation of the elastic constants of a solid containing holes. We consider the mass of the sediment replaced by a grain of conductivity $k_1$ with a film of water of conductivity $k_2$ uniformly adhering to it, the whole being surrounded by a material of conductivity $k$ which is to be determined. For the void-ratio of the material, the following relations can be easily seen to hold for the various models:

(1) $\epsilon = x_2/x_1$

if $x_1$ is the thickness of the plate (grain) and $x_2$ that of the water film.

(2) $\epsilon = r_2^2/r_1^2 - 1$

where $r_1$ is the radius of the cylindrical grain and $r_2 - r_1$ is the thickness of the cylindrical water film.

(3) $\epsilon = r_2^3/r_1^3 - 1$

where $r_1$ is the radius of the sphere surrounded by a water film of thickness $r_2 - r_1$.

(4) As for model (1).

It is easy to show that

$$\frac{k}{k_1} = \frac{\kappa + \epsilon\kappa}{\kappa + \epsilon}$$

for model (1) and

$$= \frac{1 + \epsilon\kappa}{1 + \epsilon}$$

for model (4) where

$$\kappa = \frac{k_2}{k_1}.$$

For model (3), take spherical coordinates with the origin at the centre of the grain and let the general heat flow be perpendicular to the plane $r \cos\theta = \text{const}$. By symmetry, the temperature will be independent of the $\phi$-coordinate In the solid of conductivity $k$, the temperature for unit rate of heat flow is everywhere $r \cos\theta/k$. In the shell of conductivity $k_2$ (water film) the appropriate solution of Laplace's equation for the temperature is

$$\Sigma\,(a_n r^n + b_n r^{-n-1}) P_n(\cos\theta)$$

and in the spherical grain it is

$$\Sigma\, c_n r^n P_n(\cos\theta).$$

The boundary conditions are: (i) the temperature should be continuous for $r = r_1$ and $r = r_2$; (ii) the radial heat flow should be continuous for $r = r_1$ and $r = r_2$. They lead to the following four equations:

$$\begin{aligned} r_2/k &= a_1 r_2 + b_1 r_2^{-2}, \\ c_1 r_1 &= a_1 r_1 + b_1 r_1^{-2}, \\ 1 &= k_2(a_1 - 2b_1 r_2^{-3}), \\ k_1 c_1 &= k_2(a_1 - 2b_1 r_1^{-3}). \end{aligned}$$

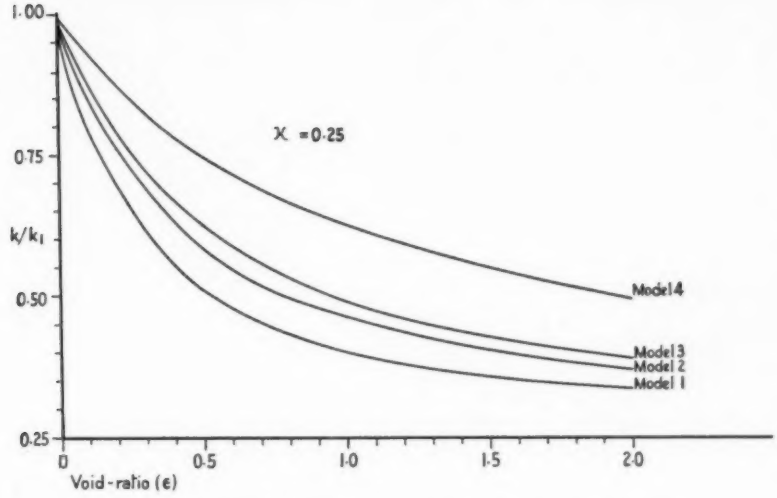


FIG. 12.—Variation of heat conductivity in a porous sediment.

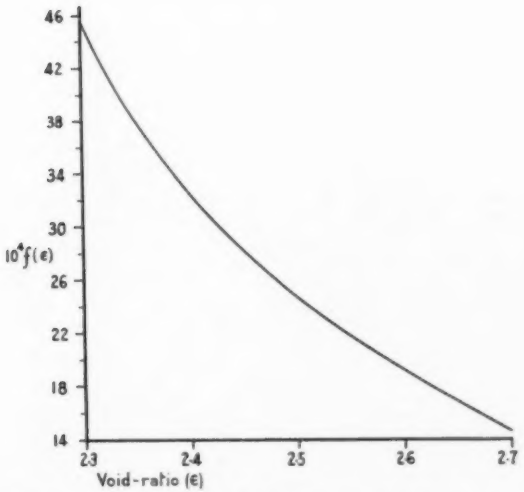


FIG. 13.—$f(\epsilon)$ for Globigerina Ooze (Sample 1).

$$f(\epsilon) = \frac{(1+\epsilon)(0{\cdot}5+1{\cdot}25\epsilon)}{0{\cdot}5+0{\cdot}3\epsilon}\exp(-\epsilon/c)$$

The last two equations can be solved for $a_1$ and $b_1$ and then substitution in the first equation gives:

$$\frac{k}{k_1} = \frac{3\kappa + \epsilon\kappa(1+2\kappa)}{3\kappa + \epsilon(1+2\kappa)}.$$

Similarly, by taking cylindrical coordinates we find that, for model (2)

$$\frac{k}{k_1} = \frac{2\kappa + \epsilon\kappa(1+\kappa)}{2\kappa + \epsilon(1+\kappa)}. \tag{21}$$

As a typical example of deep sea sediments let us take Globigerina Ooze which has a very high percentage of calcium carbonate. The conductivity of calcium carbonate ($k_1$) is $60 \times 10^{-4}$ c.g.s. units and that of water ($k_2$) is about $15 \times 10^{-4}$ c.g.s. units. Hence $\kappa = 0{\cdot}25$.

Figure 12 shows the plot of $k/k_1$ against $\epsilon$ for all the four models. The cylindrical model (No. 2, Equation 21) will be chosen now as a working hypothesis. Using the variation of $\epsilon$ with $z$ for Globigerina Ooze given in Figure 9, the variation of $k/k_1$, with $z$ is readily found.

Going a little further, we can calculate the average conductivity $k$ of a given thickness $Z$ of sediments. Obviously,

$$Z/\bar{k} = \int_0^z dz/k. \tag{22}$$

Using equation (16) we find that,

$$Z/\bar{k} = -\{p_1/k_1 C(\gamma_s - 1)\} \exp(\epsilon_1/C) \int_{\epsilon_1}^{\epsilon} f(\epsilon) d\epsilon \tag{23}$$

where, for model 2,

$$f(\epsilon) = (1+\epsilon)(k_1/k).$$

The integral in (23) can be evaluated graphically. As an example, $f(\epsilon)$ for Globigerina Ooze is plotted in Figure 13. If $Z = 10$ m the upper limit is found to be 2·35 from equation (16). On evaluating

$$\int_{2{\cdot}59}^{2{\cdot}35} f(\epsilon) d\epsilon$$

graphically and putting appropriate values for the constants in equation (23), $\bar{k}$ is found to be $40 \times 10^{-4}$ c.g.s. units.

## 7. The mechanism of the compaction of sediments

7.1. The manner in which sediments compact has no semblance to Hooke's law in elasticity inasmuch as their compression is almost completely irreversible and the "strains" are relatively large. Loose sediments are often assumed to be composed of small spherical particles. The void-ratio of an aggregrate of uniform

* The author acknowledges gratefully the help of Dr A. H. Cook in this attack on the problem.

spherical particles in contact with each other can never be zero but lies between 0·92 and 0·35. Between these two extreme cases there are six other stable arrangements (Dallavalle 1948) in each of which the void-ratio and the form of a single pore are different. A simple consideration shows that the potential energy of the spheres in the aggregate of the lowest void-ratio is smaller than in any of the other aggregates so that packings of uniform spherical particles will tend to pass to that arrangement when disturbed. If the spheres be initially dispersed in a liquid, then on the application of a small stress (with a free drainage of the liquid) they will come closer together until they touch each other. At this point the observed compression will cease and the void-ratio will lie between the above mentioned limits. We need not, *a priori*, assume any particular arrangement of the packing. If it is not in the minimum-void-ratio state, then on the application of a further stress (probably quite small) the packing will "tumble down" to that state and a "sudden" decrease of void-ratio will be observed. When this stage is reached further increase of stress will involve the deformation of the individual grains and, provided the crushing strength is not exceeded, this deformation will be reversible. When the stress exceeds the crushing strength, the voids will tend to be filled with the crushed material. Two kinds of crushings can be distinguished. If each sphere in the aggregate be cut into several fragments along the planes of great circles, the minimum void-ratio of the "new" aggregate still remains 0·35. But the crushing will, in general, be irregular and then the void-ratio may be smaller and even approach zero.

Some of the above considerations can be tested on rocks known to be compacted mechanically with very little cementation, e.g. Triassic sandstones in England. Again, if the sediments are laid down grain by grain and are then cemented, a slide of the rock can be analysed and the (original) porosity in the absence of the cement determined by adding the volume of the cement to the observed volume of the voids. Sorby (1908) analysed many slides in this way. In Table 4 some data on the observed void ratios of such "granular rocks" are given for comparison with the theoretically estimated value of 0·35 for spherical particles.

**Table 4**

*Void-ratios of "granular rocks"*

| Rock type | Observed void-ratio | Observer | Remarks |
|---|---|---|---|
| Lincolnshire Oolite | 0·32 | Sorby | Slide technique |
| Portland Oolite | 0·30 | Sorby | Slide technique |
| Oolitic beds in carb. limestone near Bristol | 0·67 | Sorby | Slide technique |
| Carb. limestone near Bristol | 0·33 | Sorby | Slide technique |
| Wenlock limestone | 0·22 | Sorby | Slide technique |
| Triassic sandstone | 0·36 | Parasnis | Direct measurement |
| Tea Green Marls | 0·35 | Parasnis | Direct measurement |
| Carbonaceous sandstone | 0·23 | Parasnis | Direct measurement |

The values for Oolites compare favourably with the theoretical value although on the lower side. This may be due to the fact oolitic grains depart rather widely from a spherical shape. Triassic sandstone and Tea Green Marls give excellent agreement with theory. In carbonaceous sandstones, organic matter probably lowers the void-ratio under the theoretical minimum.

7.2. *Sediments as chemical complexes.* In the context of compaction tests, most of the important sediments are clays exhibiting properties like thixotropy and plasticity. A clay with its water content between the plastic and the liquid limits is not merely an aggregate of mineral grains with water added to it but is subject to intermolecular forces arising from adsorbed cations, capillary forces arising due to the water and other bodily forces due to the crystalline nature of the clay minerals. The chemical constitutions of the liquid and of the grains are important factors in determining the thickness of the liquid films by which the grains will be separated and held in equilibrium. In general, all kinds of liquids which contain (OH) groups and active H atoms (e.g. water, alcohol, acetic acid) produce with clay particles a system whose properties are akin to what are usually termed "clays". Such a system cannot be obtained with hydrocarbons like petrol or benzine. When a clay–water system is subjected to an external pressure, the equilibrium of the water films is disturbed—a fact which is manifested in the expulsion of water. A clay–air system, if it consist of uniform spherical particles, will initially have a minimum void-ratio of 0·35 but a clay–water system in equilibrium can have almost any void-ratio and usually has void-ratios exceeding 1·5 under small pressures. At such large void-ratios much of the water presumably exists as "free" water except for small forces on its molecules while some is adsorbed on the grains. For very low void-ratios, almost all the water present will be adsorbed and more strongly bound than the free water of the earlier stages. Therefore a greater pressure will be necessary to expel it or in other words, for a given pressure only a small amount will be expelled which leads us to the conclusion that the compression index will decrease with decreasing void-ratio. The forces binding the water will, of course, depend on the mechanism of adsorption.

7.3.1. A question may be naturally asked whether it is possible to derive a law of compaction from the above considerations. Let us suppose that we have a sample of clay occupying a length $l$ and a volume $V_0(1+\epsilon)$ where $V_0$ is the volume of the grains and $\epsilon$ is the void-ratio. Let us further suppose that the particles of the clay are held apart by some forces and that an energy $e$ is needed to force the water out and bring two particles closer. Let $kdp$ be the number of such ruptures per cubic centimetre of the grains for an increment of the pressure from $p$ to $p+dp$. If each particle has contacts with, say, $m$ particles the energy needed to rupture the contacts in a volume $V_0$ will be $V_0mekdp$. When the water is forced out, the length of a laterally confined clay sample will decrease by $dl$ and the work done by the piston in compressing it will be $pAdl$ if $A$ is the area of cross-section of the sample. Since $Al = V_0(1+\epsilon)$ the work done can be written as $pV_0d\epsilon$. Hence,

$$-pV_0d\epsilon = V_0mekdp$$

which gives, on integration,

$$\epsilon = \epsilon_1 - mek\ln p/p_1 \tag{33}$$

where $\epsilon_1$ is the void-ratio at $p_1$.

If $p_1$ is unit pressure, (33) is Terzaghi's relation with

$$C = mek. \tag{34}$$

Equation (34) provides us with a fundamental method of finding $C$; $m$ depends on the shape of the clay particles and the lattice structure of the clay; $e$ on the nature

of the forces binding the particles. $k$ is probably typical of a given clay but may vary with pressure.

The order of magnitude of $k$ when the binding between the particles is of a purely capillary nature can be found as follows. Let us suppose that the particles are perfect spheres with an average radius $r$. Then $m = 6$ and $e = 2r^2T$ where $T$ is the surface tension of water. Consequently,

$$C = 12kr^2T.$$

Some typical values are:

$$C = 0{\cdot}30,$$
$$r = 5 \times 10^{-4} \text{ cm},$$
$$T = 80 \text{ ergs/cm}^2.$$

Hence $k$, the number of ruptures per cubic centimetre for unit increase of pressure, will be approximately $1{\cdot}2 \times 10^3$.

## 8. General conclusions

From the experiments described in this paper the following conclusions can be drawn:

(1) Departures are to be expected at high pressures from Terzaghi's linear relation between the void-ratio of a sediment and the logarithm of the applied pressure, although some sediments appear to follow the relation up to several hundred kg/cm$^2$. However, the departures are not such that large errors would arise in assuming the validity of Terzaghi's relation up to, say, 1 000 kg/cm$^2$.

(2) The void-ratio at unit pressure is proportional to the compression index which suggests that the pressure at which the state of no voids is reached is the same for all sediments. This pressure can be estimated to be of the order of 2 500 kg/cm$^2$.

(3) It is possible to find the stress to which a given clay sample has been subjected and therefore it should be possible to estimate the thicknesses eroded from the present tops of clayey sediments.

## 9. Acknowledgments

The work described above was carried out at the Department of Geodesy and Geophysics, Cambridge University, in 1950–1951 during the author's tenure of the Shell studentship. He wishes to thank the Royal Dutch-Shell Group of Oil companies for the studentship.

Thanks are also due to the authorities of the British Museum of Natural History, to the Columbia University and to the London Brick Company for kindly providing some of the samples used in the above experiments.

Finally, the author would like to thank Miss K. Thrysin and Mrs M. Tollert for help in preparing the manuscript.

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# Seismic Energy from Explosions

P. N. S. O'Brien

(Received 1959 June 16)

*Summary*

For underwater explosions it is shown that the amplitude of the first arrival of the ground wave should be proportional to $W^{\frac{1}{3}}$, where $W$ is the charge weight. Also that its frequency content should be independent of $W$. Of the total chemical energy of the explosive it is estimated that about 3 per cent is available in the frequency range 0–100 c/s.

For underground explosions it is postulated that the radiated pressure pulse begins to obey infinitesimal strain theory once its impulse per unit area has decreased below a limiting value. This value is constant for a given rock type. From this it follows that the seismic amplitude should usually be directly proportional to $W$. But for high frequencies or very large charges it should become proportional to $W^{\frac{1}{3}}$.

These predictions are compared with observational data, both original and those previously published. They agree well for underground explosions. For underwater explosions the data are themselves inconsistent and more are needed.

## Introduction

In seismic exploration a quantity of dynamite is exploded a small distance beneath the Earth's surface and a record made of the resulting vibrations of the ground. When comparing one record with another, or when predicting the charge necessary adequately to record signals reflected or refracted from particular horizons, it is often desirable to know how variation of the weight of charge fired affects the recorded signal.

This note presents a simple quantitative discussion of the effect of charge weight on seismic amplitude and frequency and compares its predictions with the few experimental data which are available.

## A. Underwater explosions

### 1. General theory

When a charge of explosive is detonated underwater the immediate result is a very large amplitude pressure pulse roughly in the form of an exponential spike. The duration of the spike increases as the cube root of the weight of charge and is a few tenths of a millisecond for a one pound charge. At regular intervals thereafter, a series of pulses of much smaller amplitude is radiated. The form of these

secondary pulses is roughly that of a double exponential spike, and each normally carries only a few per cent of the total energy in the original primary pulse. In seismic refraction shooting most reliance is placed on the first arrival, and the secondary pulses occur too late to contribute to this. On the other hand, in reflection shooting, it is the later events on the record which are important, but in this case the charge is placed at a shallow enough depth to ensure that no secondary pulses are radiated.

We are therefore going to assume that the seismic waves which are of interest are produced by the incidence of the primary pulse on the interface at the bottom of the water. The effects of reflections of the primary pulse from the surface of the water will be ignored. This effect is a straightforward interference phenomenon and not relevant to the laws relating the amplitude and shape of the observed seismic waves to the charge size. We shall also ignore the long period rarefraction which follows the initial compression pulse. This will not normally be registered by the seismic instruments, which seldom pass frequencies less than about 5 c/s.

The primary pressure pulse may be closely approximated in form by a function of time, $f(t)$, which is zero for all time before the instant of detonation, and is a monotonically decreasing function at all later times. Although it is usual to represent $f(t)$ by a simple exponential term, experiment shows that a single time constant gives a very poor approximation for pressures less than about 30 per cent of the maximum. This may be seen in Figure 1, which shows a typical plot for the primary pulse. Consequently, $f(t)$ is defined as follows,

$$\begin{aligned} f(t) &= 0 && t < 0, \\ f(t) &= P\exp(-at) && 0 \leqslant t \leqslant T, \\ f(t) &= bP\exp[-(an)(t-T)] && t \geqslant T, \end{aligned}$$

where $P$ is the maximum pressure, and $a$, $b$, $n$ and $T$ are suitable constants.

This function is shown by the dashed lines in Figure 1. The frequency spectrum of this idealized function may be found from the spectrum of a single exponential spike.

For such a spike,

$$\begin{aligned} F(t) &= 0 && t < 0, \\ F(t) &= P\exp(-at) && \infty > t \geqslant 0, \end{aligned}$$

and its Fourier transform is $P/(a+i\omega)$ where $i$ is the imaginary operator and $\omega$ is the circular frequency. The Fourier transform for the pulse $f(t)$ may therefore be written as

$$P/(a+i\omega)-bP\exp(-i\omega T)/(a+i\omega)+bP\exp(-i\omega T)/(na+i\omega).$$

The modulus of this expression gives the amplitude spectrum and the argument gives the phase spectrum.

Because of absorption, scattering and other dissipative processes the frequencies in the ground wave seldom exceed 100 cycles per second. As may be found from equation (2) below, for all likely charge weights $1/a$ is at most one millisecond, and will usually be closer to one-tenth of a millisecond. Providing all processes in forming and transmitting the ground wave are linear the principle of superposition may be used to argue that, of all the frequencies present in the shock wave, only those which are observed in the ground wave are relevant to the

present discussion. The principle of superposition also allows us to state that a proportionate variation in the amplitude of a particular frequency component in the initial pulse will produce an identical proportionate variation in the amplitude of that component in the ground wave.

$T$ is less than a millisecond, and so, because the ground wave contains only low frequencies, we may put $\cos \omega T = 1$ and $\sin \omega T = \omega T$. Also, $\omega^2$ may be neglected compared with $n^2a^2$. With these approximations the transform for $f(t)$ becomes

$$[P/a - bP/a + bP/na] - i[(\omega P/a^2)(1 - b(aT+1) + (b/n^2)(naT+1))]$$
$$= kP/a - i(g\omega P/a^2)$$

where $k$ and $g$ are constants defined by the shape of the shock wave.

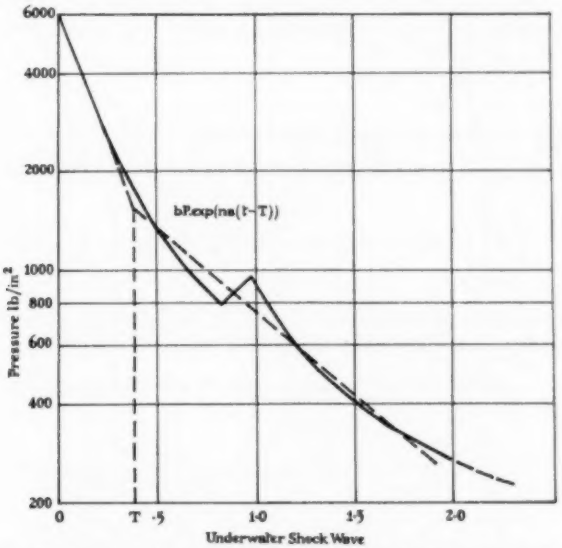


FIG. 1.—Pressure against time 14 ft from an 80 lb charge of 50/50 pentolite (Cole 1948, Figure 7.1, p. 230).

We finally obtain the low-frequency amplitude spectrum as

$$A = Pk/a$$

and the low-frequency phase spectrum as

$$X = -g\omega/ka.$$

Now the scaling laws for the initial pulse are given (Cole 1948, equation (7.6) and Figure 7.9) by the equations

$$P = s(W^{\frac{1}{3}}/R)^q \qquad (1)$$

and

$$1/a = CW^{\frac{1}{3}}(R/W^{\frac{1}{3}})^\gamma$$

where $W$ is the charge in pounds weight, $R$ is the distance in feet from the explosion and $s$, $q$ and $C$ are constants for a particular explosive*. For TNT, $q = 1{\cdot}13$,

* 1 ft = 30·40 cm; 1 nm = 1·853 km; 1 lb = 453·6 g; 1 psi = $6{\cdot}89 \times 10^4$ dyn/cm².

$r$ is about $\frac{1}{4}$ and $C$ is about 0·05 milliseconds/(pound weight)$^{\frac{1}{4}}$. Using these values for $q$ and $r$ we obtain the variation with charge weight of the spectral components present in the ground wave as,

$$A = YW^{1\cdot13/3}W^{3/12} = YW^{0\cdot63} \qquad \text{for } \omega \ll a,$$

and

$$X = ZW^{\frac{1}{4}}\omega \qquad \text{for } \omega \ll a,$$

where $Y$ and $Z$ are constants.

These equations show that there should be no appreciable change of shape of the ground wave with variation in charge weight, and that its amplitude should vary as $W^{0\cdot63}$. This value is for TNT; for other explosives it may be slightly different.

It has always been somewhat of a puzzle that variations in charge weight produced no change in the form of the ground wave. The above reasoning suggests a solution.

## 2. Percentage energy in the ground wave

The total energy in the initial pressure pulse is given approximately by

$$\frac{S}{Z}\left[\int_0^T [P\exp(-at)]^2\,dt + \int_T^\infty [bP\exp(-na(t-T))]^2\,dt\right]$$

where $Z$ is the acoustic impedance of the water, and $S$ is the area of its wavefront.

Integrating between the limits, we obtain the total energy as

$$(P^2S/2aZ)\,.\,[1-\exp(-2aT)] + b^2P^2S/2naZ.$$

Now,

$$T \sim 1/a,$$

$$n \sim 1/3, \text{ and}$$

$$b \sim 1/e,$$

therefore the total energy is $SdP^2/2aZ$, where $d$ is not very different from unity.

The energy in the low-frequency band is given by

$$(2S/Z)\int_0^F A^2\,df = \int_0^F (2Sk^2P^2/Za^2)\,df = 2Sk^2P^2F/Za^2 \qquad \text{for } 2\pi F \ll a.$$

Therefore, the percentage energy in the ground wave is given approximately by $(2k^2P^2F/a^2)(2a/P^2)\times 100 = 400k^2F/a$, where $F$ is the highest observed frequency in the ground wave spectrum. For a 5 lb charge of TNT, $k$ is about 1·2 and $a$ is about $10^4$. These values vary slightly with $R$, but ignoring this we may calculate the useful seismic energy as about $600\times 10^{-4}F$ per cent of the energy in the initial pulse. At one charge radius* the initial pulse contains about one-half of the total energy released by the explosion (Cole 1948, p. 146).

Because the ground wave seldom contains frequencies higher than about 100 cycles per second, only some 3 per cent of the explosive energy is useful for seismic exploration. Of this amount a portion will be reflected at the bottom of the water and therefore lost to the ground wave.

* Equals (weight of charge in pounds)$^{1/3}$ feet.

## 3. Experimental results for underwater explosions

(*a*) *Amplitude and pulse shape.*—An examination of the relevant literature has provided only one series of quantitative data on the variation of the amplitude of the ground wave with the weight of charge fired. This is by Ewing (1948, Table 5, p. 22) and it is given in Table 1. Although he states that it follows a $W^{\frac{1}{3}}$ law, a least squares fit gives $W^{0.43\pm0.015}$. This is for charges from $\frac{1}{3}$ lb to 300 lb of TNT. The index is appreciably, and significantly, lower than the above theory predicts. He also states, in agreement with all other observers, that there is no change in pulse shape for this same range of charge weights.

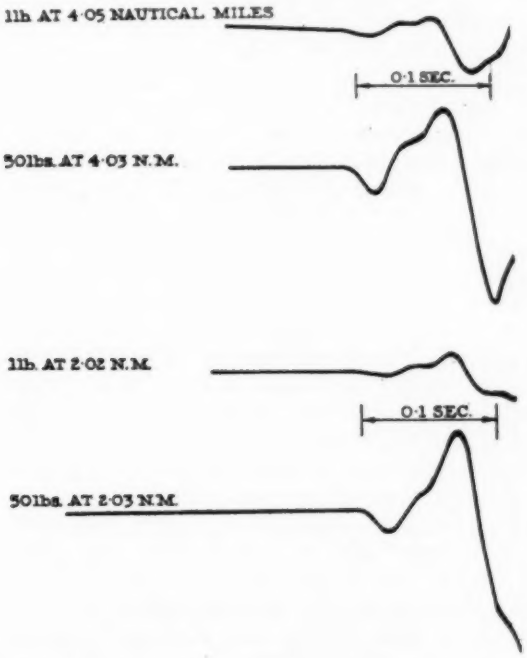


FIG. 2.—Hydrophone records for different sized shots.

Another series of measurements has been taken off records loaned to the British Petroleum Company by the Admiralty Research Laboratory, Teddington. This series is for charges of 1 lb and 50 lb. The first arrivals for the 1 lb charges were too weak for their "periods" to be read accurately. But, for the same distance, the difference between the "period" for a 1 lb charge and that for a 50 lb charge was certainly no greater than 20 per cent, and was probably zero. This is illustrated by the tracings shown in Figure 2. The amplitude measurements are shown in Table 2; a least squares fit gives amplitude proportional to $W^{0.67\pm0.02}$.

The observations of Worzel and Ewing were made for shots fired both on and off the sea bottom and in various depths of water. The A.R.L. charges were all laid on the sea bottom in about 20 fathoms of water.

(*b*) *Available seismic energy.*—No experimental estimate of the total energy in the ground wave is available for underwater explosions, though it is known that

## Table 1

*Table 5 of Worzel and Ewing* (1948)

*Intensity difference between various charges and a 25 lb charge*

Number of decibels to be added to the intensity from a charge as given at the head of the column to obtain the intensity which would have been obtained from a 25 lb charge.

| | Trace 2 | | | | | Trace 3 | | | | | Trace 4 | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | 300 | 5 | 4·5 | 3·5 | 0·5 | 300 | 5 | 4·5 | 3·5 | 0·5 | 300 | 5 | 4·5 | 3·5 | 0·5 |
| Solomons Shoal | | | | | | | | | | | −5 | +7 | | | +16 |
| Solomons Deep | | | | | | | | | | | | | | | |
| Jacksonville Shoal | | | | | | | +6 | | | +15 | | | | | |
| Jacksonville Deep | | | | | | −8 | +6 | | | +14 | −9 | | +6 | | +14 |
| Virgin Islands Shoal | | | | | | −7 | | | | | −7 | | +8 | | +18 |
| Virgin Islands Deep | | | | | | | | | | | −8 | | | | |
| Barbados | −7 | | +7 | | +16 | | | | | | −9 | | +6 | | +14 |
| Orinoco Shoal | −8 | | | +5 | +10 | −11 | | | +6 | | −7 | | +6 | | +17 |
| Orinoco Deep | −7 | | | +8 | +15 | | | | | | −10 | | +10 | | +20 |

## Table 2

*Amplitude ratios for 50 lb and 1 lb charges from the A.R.L. records*

| Distance in nautical miles | 2 | 2 | 2 | 2 | 4 | 4 | 4 | 4 | 6 | 6 | 6 | 6 | 8 | 10 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Amplitude ratio | 12·0 | 15·4 | 11·3 | 11·3 | 10·0 | 15·4 | 10·5 | 14·3 | 13·0 | 14·8 | 13·3 | 16·5 | 15·0 | 18·0 |

far smaller charges are required for refraction shooting at sea than on land. However, comparison of shots made in boreholes with those fired in reservoirs suggests that underwater explosions are only about half as efficient (Gamburtsev & others 1950) so the lower charges needed when exploring the ocean would appear to point to simpler geological conditions.

## B. Underground explosions

### 1. General theory

The mechanical processes involved when a charge of dynamite is detonated in a well-tamped borehole have been aptly described by Morris (1950a). If the rock is hard and compact, such as sandstone or limestone, then the material immediately surrounding the charge is crushed and shattered. Further out from the walls of the shot-hole the pressures are reduced sufficiently to prevent shattering but are still high enough for non-linear processes to apply and for the peak pressure to be reduced with distance at an appreciably faster rate than for a simple acoustic wave. For more plastic materials such as clays, marls and shales much less crushing and shattering is observed and, instead, a larger cavity is created by the material being forced outwards and increasing the density of the surrounding rock. In any event, the disturbance is soon reduced to a level at which infinitesimal strain theory may reliably be applied. The surface separating an inner volume in which the stresses are too great for infinitesimal strain theory to hold from the outer volume in which it may be applied is often referred to as the "equivalent radiator" (e.g. Sharpe 1946; Gaskell 1956). For a given medium the limiting condition at the surface of the radiator is a constant and equals that at which infinitesimal theory may just be applied. A larger charge will form a larger equivalent radiator but, for a given medium, will not give different limiting conditions from those due to a smaller charge. As a limiting condition we suppose that the impulse per unit area is constant. The equivalent radiator is a somewhat nebulous concept whose surface has not been observed. Even if measurements were made to locate it they could not be expected to define it very precisely, as the transition between the region where infinitesimal theory may not be applied to the region where it may be applied probably will not be sharp. Nevertheless, it seems reasonable to suppose that its volume is directly proportional to the volume of the cavity formed by the explosion. Also, the well-known scaling laws for explosions (e.g. Lampson 1946; Gaskell 1956) provide that the volume of the cavity is directly proportional to the charge weight, so the volume of the radiator will also be directly proportional to weight of charge.

If we assume the radiating surface to be spherical and the pressure over it to be uniform then it has been shown by Blake (1952) that the amplitude $A$, of the particle velocity is given by

$$A = \frac{PK}{\rho c^2} \cdot \omega \cdot [(\omega/c)^2+(1/r)^2]^{\frac{1}{2}} \cdot (a^3/r)[1+(1-2K)(\omega a/c)^2+K^2(\omega a/c)^4]^{-\frac{1}{2}}$$

where, $P$ = amplitude of applied pressure,
$K = (1-\sigma)/2(1-2\sigma)$,
$\sigma$ = Poisson's ratio,
$\rho$ = density,

$c$ = dilatational wave velocity,
$\omega$ = angular frequency,
$a$ = radius of equivalent radiator,
$r$ = distance from centre of equivalent radiator.

The duration of the pressure pulse close to the shot is very short, and largely unidirectional, so that a similar analysis to that just presented for an underwater explosion gives the low-frequency amplitude spectrum as directly proportional to $P_m\alpha$, where $P_m$ is the peak pressure and $\alpha$ determines the duration of the transient.

Impulse per unit area is the time integral of the pressure so, providing the shape of the pressure pulse does not vary much over the short distances involved, constancy of impulse implies constancy of the low-frequency spectrum of the pulse. The amplitude of the particle velocity at distances such that $(\omega r/c)$ is much greater than unity is then proportional to $f(x)$, $x$ equals $\omega a/c$, where

$$f(x) = x^3[1+(1-2K)x^2+K^2x^4]^{-\frac{1}{4}}.$$

This function has been graphed in Figure 3 for typical values of $\sigma$.

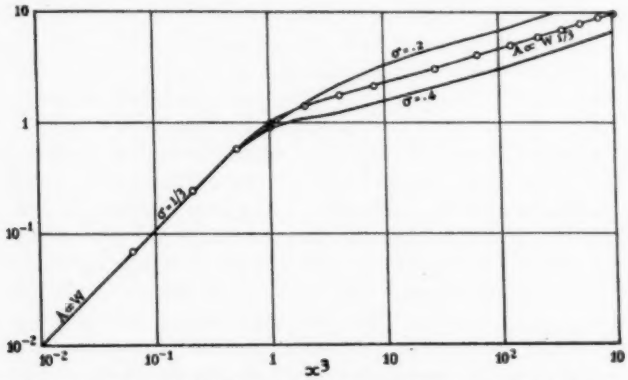


FIG. 3.—$f(x)$ as a function of $x^3$. $f(x) = x^3(1+(1-2K)x^3+K^2x^4)$.
$W$ = Weight of charge.
$\sigma$ = Poisson's ratio.
$A$ = Particle velocity.

In seismic exploration $\omega$ is seldom greater than 500 $s^{-1}$, and $c$—at the shot—is not often less than 5 000 ft/s. The radius $a$ is not well specified but the measured values for peak pressure and radial strain quoted in Tables 3 and 4 would

**Table 3**

*Peak pressure in clay* (*Lampson* 1946)

| Charge weight lb | Distance ft | Peak pressure psi |
|---|---|---|
| 8 | 8·2 | 40 |
| 216 | 15·5 | 360 |
| 1 000 | 26 | 300 |

**Table 4**

*Peak radial strain for different rocks (Duvall & Atchison 1957)*

| Charge weight lb | Distance ft | Peak radial strain | Rock type |
|---|---|---|---|
| 3·9 | 10 | 0·0005 | Granite |
| 24 | 10 | 0·0008 | Marlstone |
| 8 | 10 | 0·0006 | Sandstone |
| 16 | 10 | 0·0014 | Chalk |

indicate that for charges less than about 100 lb it is not likely to be as much as 10 ft. For most rocks $K$ will not be too different from unity so that for the majority of seismic shots $x$ becomes less than unity and therefore $f(x)$ differs little from $x^3$. The amplitude $A$, then becomes proportional to $a^3$ and, therefore, to the weight of charge fired.

This will be true for all the frequencies in the observed pulse and therefore there will be no change with charge weight in the shape of the recorded pulse.

In long-distance refraction shooting, where charges of several thousands of pounds of dynamite are quite common, and in near-surface reflection shooting, where frequencies of three or four hundred cycles per second are recorded, $x$ will no longer be less than unity. For large values of $x$, $A$ will increase with the cube root of the charge weight and also the recorded pulse shape will lengthen slightly as the charge is increased. This may explain the well-known observation in refraction shooting, that once the charge has reached a few hundreds of pounds weight, then a very great increase is needed in order to obtain a relatively small increase in seismic amplitude.

## 2. Experimental results for underground explosions

A fair number of reports are available on measurements made close to shots fired in quarry blasting. Thoenen & Widnes (1942) and Morris (1950b) produce evidence for the relationship

$$\text{Amplitude} \propto (\text{weight of charge})^1.$$

However, insufficient data were presented in those reports to estimate a value for $x$, so no quantitative comparison with the foregoing hypothesis is possible; although, as they were measuring close to the source, it is very likely that they had an appreciable high-frequency content, with the consequence that $x$ would be greater than unity. Therefore, one would expect the amplitude of the ground motion to be proportional to some fractional power of the charge weight. More recent observations on quarry blasts, reported on by Habberjam & Whetton (1952), when fitted by least squares to the equation

$$A = KW^n,$$

yield a value for $n$ of $0{\cdot}88 \pm 0{\cdot}05$. For these experiments $x$ may be estimated as being fairly close to unity.

These results for quarry blasts are consistent with the hypothesis advanced in this paper. This may be fortuitous because the mechanism for the production of

seismic waves from very shallow shots may be different from the mechanism involved in the much deeper shots normally used in seismic prospecting.

Observations more closely related to seismic shooting have been reported on by Blair & Duvall (1954). For shots 20 ft deep in shale they found that for charges up to about 100 lb and for distances between 25 and 1 000 ft, the form of the received pulse varied with charge weight. A least squares calculation using the data in their Table 3 gives a value for $n$ of $0{\cdot}730 \pm 0{\cdot}07$. Once again these measurements were made rather close to the source so that $x$ may be greater than one.

The only other published data that can be found are by Gaskell (1956). He fired a series of charges in neighbouring holes at a depth of 120 ft and recorded the refracted waves 18 000 ft away. At this distance the predominant frequencies are 20–30 c/s. His data yield a value for $n$ of $1{\cdot}11 \pm 0{\cdot}07$.

A number of records have been examined from a routine refraction survey over the Eakring oilfield in the Midlands of England from which was obtained a value for $n$ of $1{\cdot}04 \pm 0{\cdot}08$. This value is the mean of two values obtained by different methods. The first method was the straightforward one of comparing repeat shots. There were 25 examples when two shots were exploded sequentially at the same shot-point (not necessarily the same shot-hole) and were recorded by the same geophone. It was therefore possible to plot an amplitude ratio against a charge weight ratio. The amplitude ratios were taken for each well-defined event on the record. There were usually two and sometimes three such events and their ratios agreed very closely with one another. A mean value for each record is given in Table 5. No difference could be found between pairs of shots fired in new

**Table 5**

*Data from repeat shots*

| Smallest charge lb | Charge weight ratio | Amplitude ratio |
|---|---|---|
| 20 | 1·00 | 1·26 |
| 20 | 3·00 | 2·86 |
| 20 | 1·00 | 0·83 |
| 20 | 3·00 | 4·79 |
| 40 | 2·50 | 2·13 |
| 60 | 1·67 | 1·30 |
| 60 | 1·67 | 1·47 |
| 60 | 1·67 | 1·41 |
| 60 | 1·67 | 1·02 |
| 80 | 1·25 | 0·70 |
| 80 | 1·25 | 1·52 |
| 80 | 1·88 | 2·60 |
| 100 | 1·50 | 1·80 |
| 100 | 1·50 | 1·18 |
| 100 | 1·50 | 1·53 |
| 100 | 1·50 | 1·70 |
| 100 | 2·50 | 1·55 |
| 150 | 1·34 | 1·63 |
| 150 | 1·33 | 2·13 |
| 150 | 1·00 | 0·72 |
| 150 | 2·00 | 1·76 |
| 200 | 1·50 | 1·22 |
| 200 | 1·10 | 1·05 |
| 250 | 1·20 | 0·88 |
| 300 | 1·00 | 0·82 |

holes and pairs fired in previously used holes. Therefore all the data were treated as belonging to a single group. They yield a value for $n$ of $0{\cdot}92 \pm 0{\cdot}106$.

The second method was less direct: it consisted of measuring on a number of records the first peak amplitudes associated with a particular refractor and reducing these amplitudes to a common distance by means of the known attenuation coefficient for the refractor. This was done for the two refractors which were well recorded as first arrivals. For each seismogram mean values were calculated for the amplitudes and distances. Normally, each seismogram recorded 24 geophones so that each amplitude shown in Tables 6 and 7 was obtained from 24 measurements. Values for $n$ from the two refractors were $1{\cdot}39 \pm 0{\cdot}19$ and $1{\cdot}04 \pm 0{\cdot}15$.

From the charge weights and recorded frequencies it seems certain that $x$ was less than unity both for Gaskell's experiments and for those described above.

All the values for $n$ are collected together in Table 8. They are all consistent with the hypothesis advanced here except, perhaps, for that obtained from Duvall's data.

A simple weighted mean for all the values is $0{\cdot}93 \pm 0{\cdot}033$. If Duvall's data are excluded the value becomes $0{\cdot}98 \pm 0{\cdot}037$.

In addition to his series of repeat shots in sandstone Gaskell (1956) reported the results of another interesting set of experiments. He fired a series of small charges in water-filled cavities in clay which varied in diameter between 1 in. and 14 inches. The amplitude of the first arrivals were measured with a flat response (1–100 c/s) velocity instrument 100 ft away. The data he obtained are shown in Table 9 and were originally presented in his Figure 2. He also measured the cavities made by a number of different sized charges and found that the cavity produced by a 1 g charge had a radius of about 2·25 in.

**Table 6**

*Data from first peaks of an arrival refracted from a layer about 3 000 ft deep*

| Charge weight lb | Amplitude at 20 000 ft $\mu$V |
|---|---|
| 45 | 17 |
| 50 | 15 |
| 80 | 15 |
| 100 | 28 |
| 150 | 39 |
| 150 | 40 |
| 150 | 58 |
| 150 | 68 |
| 150 | 112 |
| 200 | 82 |
| 200 | 127 |
| 200 | 212 |
| 220 | 87 |
| 220 | 110 |
| 250 | 176 |
| 300 | 133 |
| 300 | 136 |

## Table 7

*Data from first peaks of arrival from a refractor 100 ft deep*

| Charge weight lb | Amplitude at 1 500 ft $\mu$V |
|---|---|
| 2·5 | 47 |
| 2·5 | 58 |
| 20 | 406 |
| 20 | 218 |
| 20 | 1 090 |
| 20 | 228 |
| 20 | 563 |
| 20 | 769 |
| 20 | 372 |
| 40 | 900 |
| 50 | 590 |
| 100 | 3 170 |

## Table 8

*Values for n in the equation* $A = KW^n$

| Range of charge lb | $n$ | s.d. of $n$ | Number of observations | Source |
|---|---|---|---|---|
| 20–300 | 0·92 | 0·106 | 25 | This paper |
| 45–300 | 1·39 | 0·186 | 17 | This paper |
| 2·5–100 | 1·04 | 0·145 | 13 | This paper |
| 20–200 | 1·11 | 0·067 | 4 | Gaskell 1956 |
| 4–198 | 0·88 | 0·053 | 45 | Habberjam & Whetton 1952 |
| 6·25–100 | 0·73 | 0·070 | 28 | Blair & Duvall 1954 |

## Table 9

*Amplitude from shots in cavities (Gaskell 1956)*

| Cavity dia. in. | Weight g | Particle velocity cm/s $\times 10^{-3}$ |
|---|---|---|
| 1 | 0·85 | 0·3 |
| 1 | 3·5 | 1·4 |
| 1 | 45·0 | 14·0 |
| 5 | 0·85 | 1·0 |
| 5 | 1·1 | 1·0 |
| 5 | 1·3 | 1·0 |
| 5 | 2·0 | 1·4 |
| 5 | 45·0 | 16·0 |
| 8 | 0·85 | 1·4 |
| 8 | 1·1 | 1·4 |
| 8 | 1·3 | 1·3 |
| 8 | 2·0 | 2·0 |
| 8 | 3·5 | 3·2 |
| 14 | 0·85 | 2·0 |
| 14 | 1·1 | 2·3 |
| 14 | 1·3 | 2·4 |
| 14 | 2·0 | 2·3 |
| 14 | 3·5 | 5·4 |
| 14 | 45·0 | 18·0 |

If $D$ represents the radius of the prepared cavity, $r$ the radius of the equivalent radiator, and $W$ the charge weight, his data may be grouped in the following three ways.

*Case 1: $D \ll r$, $W$ varying*

For this case the argument developed above may be applied directly and for the values involved $x$ will be much less than unity so that we would expect $n$ to equal one. In fact, for the 1 in. hole $n = 0{\cdot}96 \pm 0{\cdot}05$.

*Case 2: $W$ constant, $D$ varying*

For $D$ just greater than the size of the explosive we would expect no variation of seismic amplitude with variation in $D$. If $D$ is appreciably larger than the explosive but still small enough for the impulse at the cavity wall—which will be greater than that in the incident wave because of the increase in acoustic impedance across the boundary—to exceed the limiting condition, we may find $r$ from the equation

$$C(K/D)(D/r)^m = \text{constant}.$$

The first bracket gives the variation with distance in water of the shock wave impulse (Cole, p. 242, Table 73), whilst the second bracket gives the variation of the impulse in the solid. It is to be expected that $m$ will be greater than one as the peak pressure is reduced vary rapidly with distance. For soils Lampson (1946) reports that $m$ is as much as 2·5. $C$ is a partition coefficient; it might depend on $D$, but it would not be expected to vary much for the values involved.

On the equivalent radiator hypothesis the amplitude of the particle velocity is proportional to $r^3$ and so is proportional to $D^q$ where $q = 3(1 - 1/m)$.

From the largest of Gaskell's charges (45 g) $q = 0{\cdot}09 \pm 0{\cdot}02$ while for the smallest (0·85 g) $q = 0{\cdot}73 \pm 0{\cdot}03$. The latter value for $q$ may be used to estimate $m$, which would have to equal 1·3 if the hypothesis is valid. This seems quite a reasonable value for the heavy clay involved.

It seems unlikely that the argument advanced in this case will hold when $D$ is less than about one charge radius. So that springing the hole with an initial smaller charge will not appreciably increase the seismic amplitude due to the main charge, because the size of the sprung hole will always be less than one charge radius.

*Case 3: $D \gg r$, $D$ constant, $W$ varying*

For this case the impulse at the cavity walls would be below the limiting value. The amplitude of seismic motion should then vary in the same manner as for an underwater explosion, that is it should be proportional to $W^{\frac{1}{3}}$.

Gaskell's data for his largest cavity (14 in. diameter) are rather scattered. They yield a value for $n$ of $0{\cdot}57 \pm 0{\cdot}12$. The largest charge in this series of experiments was 45 g, and it is quite likely that for this shot the impulse at the cavity wall exceeded the limiting value. The next largest charge was only 3.5 g. If the 45 g charge is omitted the value for $n$ is $0{\cdot}62 \pm 0{\cdot}3$. Although the mean value is nearer to 0·67 this is really not significant because of the large increase in its standard error.

It seems, then, that Gaskell's data for small charges are quite consistent with the hypotheses advanced for both underwater and underground explosions.

*Seismic energy from underground explosions.*—The preceding hypothesis gives no indication of the amount of energy to be expected in the seismic disturbance.

And no measurements sufficiently close to the shot have been made which would enable the energy to be computed directly.

Van Melle (1954) has estimated the energy in a pulse observed by Ricker (1951a) at about 720 ft from the explosion and found that it contained about 0·5 per cent of the initial chemical energy available. This was for a pulse with a predominant frequency of some 175 c/s. Van Melle also stated that this value was about the same as that contained in pulses he had observed at similar distances from the shot. In another paper Ricker (1951b) gave some typical spectral diagrams for his observed pulses. Recently McDonal & others (1958) have measured the absorption of seismic waves in the same formation as that used by Ricker and, accepting their value of 0·8 dB per wavelength, we may compute the energy released in the frequency range 0–175 c/s as about 2–3 per cent of the initial chemical energy available. Because of uncertainties in the attenuation factor this estimate is quite likely to be in error by a factor of two.

The energy spectrum falls off fairly rapidly with decrease in frequency so that the energy below 100 c/s is probably much less than 1 per cent of the chemical energy.

Howell and Budenstein (1955) have also reported on the energy content of pulses from explosions. From their Table 1, the energy of the primary pulse in the frequency range 6–120 c/s, 14 ft from the explosion may be computed as 4·5 per cent of the chemical energy. Because of the heterogeneity of the very near surface, where their detectors were planted, this estimate may also be in error by a considerable factor.

## C. Summary and conclusions

It is well established that variation in the weight of charge fired has no effect on the spectral content of the observed ground-wave pulse. This is true for both underwater and underground shots. Because a seismic record is essentially an interference phenomenon it might be that any small variation in the spectrum of the source pulse is obscured by the frequency selective properties of the ground. However, interference effects would not impose a constant waveform until later on in the record whereas all the measurements described above were made on the very first arrival. And also, as the recorded signals from charge weight ratios of 50:1 are identical in form, it seems simpler to assume that the emitted frequencies remain constant as the weight of charge varies. This is in agreement with the hypotheses and calculations made in this paper.

There are not enough data available either to prove or to disprove the prediction made regarding amplitude variations. For underwater explosions the measurements made on some recent A.R.L. records agree well with the prediction that amplitude should be proportional to (weight of charge)$^{\frac{1}{3}}$ but the more extensive observations of Worzel & Ewing do not.

For underground explosions all the observations are consistent with the predictions made in this note except, perhaps, those made by Blair & Duvall. But, from their series of experiments, no firm comparison is possible because it is not possible properly to estimate the frequencies involved.

It is predicted that for underwater explosions about 3 per cent of the chemical energy in the explosive is available in the range of seismic frequencies 0–100 c/s.

Experiment indicates that in underground explosions the energy in the low frequencies is certainly no more than this and may be very much less.

The common observation that much smaller charges may be used when doing refraction shooting at sea might imply that underwater explosions produce the seismic frequencies more efficiently than do underground explosions. However, for refraction work on land it has been reported (Gamburtsev & others 1950) that nearly twice as much explosive is needed when shooting in reservoirs as is required for shots in boreholes. This seems to imply that the much lower charge needed at sea is due entirely to the relative simplicity of submarine geology.

**Acknowledgments**

Many thanks are due to the Admiralty Research Laboratory for their ready agreement to the loan of their records. Thanks are also due to their Mr Weston who first suggested to the writer that the cube root scaling law was theoretically incorrect for the ground wave produced by underwater explosions. The writer is indebted to the Chairman of the British Petroleum Company for permission to publish this paper.

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*Exploration Division,*
*Chertsey Road,*
*Sunbury-on-Thames, Mx:*
1959 *June.*

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# The Remanent Magnetization of the Basic Tertiary Igneous Rocks of Skye, Inverness-shire

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(Received 1959 August 17)

*Summary*

The remanent magnetization of the basic igneous rocks of Skye, the hypersthene gabbro of Ardnamurchan, and the layered gabbro of Rhum has been examined. All the rocks possessing stable magnetization were found to be reversely magnetized. This is believed to be due to the Earth's magnetic field being reversed during the half million years or so during which the rocks were formed.

Reasons are given for supposing that the dips of the Skye rocks are formational. The ancient pole position during part of the Lower Miocene is estimated to be 70°·8 (±3°·8) N. Lat., and 159°·6 (±5°·1) E. Long., a position which is significantly different from that of the present geographic pole. It is concluded that there has been relative movement between North-west Europe and the pole since the Lower Miocene.

The directions of magnetization are used in an attempt to correlate some of the lavas and dykes of Skye. It is also tentatively suggested that the rocks of Skye, Rhum and Ardnamurchan are of approximately the same age.

## 1. Introduction

The polar paths deduced from the magnetization directions of rocks, ranging in age from pre-Cambrian to Recent, from different continents have a generally similar form (Irving 1957). The similarity in form of the paths from each continent suggests that polar wandering has occurred, and the fact that the polar path for each continent is different suggests that there has also been some relative movement between the continents. In particular, the separation of the paths for North America and North-west Europe decreased from the Mesozoic onwards and eventually converged some time in the Tertiary. This convergence may be due to the ceasing of relative movement between the two continents. It may also be due to convergence of the pole positions towards the present geographic pole which results in separations in the Upper Tertiary smaller than the errors of individual determinations. To decide if and when relative movement between the two continents ceased it is necessary to have a series of precise pole positions from rocks from as many horizons as possible between the lowest and uppermost Tertiary from both sides of the Atlantic. There are, however, several difficulties in doing this at the present time.

**Table 1**

*Tertiary pole positions estimated from the magnetization directions of North-west European rocks*

Northern hemisphere polar coordinates are given; *dp* and *dm* are the semi-axes of the elliptical error area round the pole at a probability of 95 per cent, *dp* in the colatitude direction and *dm* perpendicular to it.

| Age | Formation | Pole positions Lat. | Long. | *dp* | *dm* | References |
|---|---|---|---|---|---|---|
| Plio-pleistocene | Lavas of Central France | 77·9 N | 153·3 E | 15·7 | 20·3 | Roche 1951 |
| Plio-pleistocene | Lavas of South and South-west Iceland | 87·6 N | 146·8 E | 11·5 | 12·6 | Hospers 1953. |
| Mio-pliocene | Lavas of Central France | 73·1 N | 167·3 W | 13·6 | 19·2 | Roche 1951. |
| "Miocene" | Lavas of North and North-west Iceland | 88·6 N | 5·2 E | 9·6 | 10·3 | Hospers 1953. |
| Oligocene | Intrusive of Central France | 73·0 N | 119·0 E | 12·0 | 16·0 | Roche 1950(a) and (b) |
| "Oligocene" | Dykes of North-east England | 86·0 N | 76·0 W | 21·0 | 25·0 | Bruckshaw & Robertson 1949. |
| "Eocene" | Mull lavas | 76·0 N | 162·0 W | 2·5 | 7·6 | Dutt 1955. |
| "Eocene" | Mull intrusives | 72·0 N | 139·0 E | 11·0 | 14·0 | Vincenz 1954. |
| "Eocene" | Arran dykes | 79·0 N | 154·0 E | 12·0 | 15·0 | Leng 1955. |
| "Eocene" | Antrim lavas | 74·0 N | 133·0 E | 6·0 | 8·0 | Hospers & Charlesworth 1954. |
| "Eocene" | Lundy dykes | 73·6 N | 123·5 E | 7·8 | 10·5 | Blundell 1957. |

In Table 1 are given the pole positions and half angles of the oval curves of confidence at $P = 5$ per cent (Fisher 1953, Irving 1956) obtained by earlier workers on rocks of Tertiary age from North-west Europe. It is not possible to arrange the results into any sort of order for the following reasons.

(a) All the rocks studied are igneous, and their dating has so far been done on the scanty fossil content of associated sediments. In the British Isles, Greenland and Iceland, the flora have been used and their interpretation has varied. Thus, the plants from Ardtun in Mull were first studied by Forbes (1851) and dated as Miocene. W. H. Baily (1869) came to the conclusion that the beds were of Lower Miocene age and thought they were the same age as the plants from the inter-basaltic horizon of Ulster in North-east Ireland and the Atanekerdluk formation of East Greenland. Heer (1868) thought that the plants from among the lavas of North and North-west Iceland, Greenland and the British Isles were of Miocene age. Later, however, J. Starkie Gardner (1885a, b) made a more complete study of the plants, and the following is the possible sequence deduced by him:

North and North-west Iceland—Much later than Mull, Ulster and Greenland. Could be as late as Miocene.

Lough Neagh and Bovey Tracey—Covered a long time. Probably Lower Eocene to Middle Miocene.

Ardtun—Younger than Ulster and Greenland, but still Lower Eocene.

Greenland—Younger than Ulster.

Ulster—Equivalent to the Heersian of Gelinden, i.e. Lower Eocene.

Seward and Holtum (1930) agree with the Eocene age. Hawkes (1938) expressed the view that the Iceland lavas were Eocene or possibly Oligocene in age and argued that this did not necessarily conflict with the fossil evidence.

More recently J. B. Simpson (1951) concluded from pollen studies that the bulk of the Mull lavas are most probably of late Miocene age; in a private communication he has subsequently stated that he thinks it probable that the Skye lavas sampled during the magnetic investigations are also of Miocene age. Starkie Gardner's sequence is not necessarily invalidated by this, but it seems that the beds are all younger than he supposed. However, the situation is not satisfactory.

The Lundy dykes have been stated to be pre-Bovey Tracey beds, i.e. pre-Upper Oligocene in age (Reid 1913), and Blundell (1957) concluded from the similarity of the mean magnetization direction of the dykes with that of the Antrim lavas that they were Eocene. In view of the doubt of the age of the Bovey Tracey beds and the Antrim lavas, this conclusion may have to be revised.

The intrusives of Mull and the dykes of Arran and North-east England are all evidently later than the Mull lavas, but their relative ages are still open to doubt.

The lavas and intrusives of Central France are well dated by means of plants, mammals and insects (Gignoux 1955).

(b) The estimated pole positions, except possibly those from the lavas of Mull and Antrim, are not accurately defined and are therefore of limited use for studies of polar wandering and continental drift.

(c) It is often very difficult to determine the amount by which igneous rocks have been tilted subsequent to their formation. Lavas are not always laid down horizontally, and when no sediments, half-filled vesicles, or other means of finding out their depositional attitude are available, it is not possible to determine the amount of post-formational tilting which has occurred. It is quite possible that some of the pole positions are in error for this reason. Thus one would expect the positions estimated from the lavas of Antrim and Mull to be closer to each other in view of their similar age. The Antrim lavas were corrected for dip (1° or 2°) while in the case of Mull dip corrections were not applied or even discussed, though dips of the order of 20° do occur.

(d) The secular variation has not been averaged out in some instances. It is likely that in a succession of lava flows, or a large intrusive body which has been adequately sampled, the effect of secular variation will have been averaged out. This may not be so in the case of the dykes which have been studied, as the time taken for their formation could be quite small compared with the period of secular variation.

It is quite clear that there is need for much more work on Tertiary rocks. A study of Skye was undertaken for the following reasons:

(i) To provide further data on the position of the pole during the Tertiary. Skye and Greenland are the two outstanding parts of the Thulean igneous province with large volumes of lava whose magnetization directions have not been studied.

(ii) To provide more information regarding reversals of the Earth's magnetic field. If, as some workers feel, reversals of the field occurred at regular intervals during the Tertiary, then normally and reversely magnetized rocks should be roughly equally abundant. As much information as possible is needed to decide whether or not this is the case.

(iii) To see if any information on the Tertiary geology of the island could be provided (see Section 2).

For comparison with the gabbros of Skye, it was decided to investigate gabbros from Rhum and Ardnamurchan. The magnetic properties of gabbros have not been as extensively studied as those of basalts and red sandstones. It is felt that gabbros could give useful magnetic information (Green & Jaeger 1956) and that further information of their magnetic properties would be useful.

## 2. Geology and sampling

(a) *Skye.* The Tertiary geology of Skye is shown diagrammatically in Figure 1 on which the sampling sites are also indicated. A series of lavas with individual flows ranging in thickness from 6 to 50 ft occupy most of the North-western half of the island. They are believed to form five groups of lavas, whose relationships are probably as shown in Figure 1(b) (Geological Survey, private communication). Groups 1 and 2 may be of the same age. Three orientated samples were taken from the generally massive central part of each lava from the localities shown in Table 2.

**Table 2**

*The lavas of Skye* (See Figures 1(a) and 1(b))

| Group | Rock types | Sampling locality | Ref. to Fig. 1 | No. of flows present | No. of flows sampled | Remarks |
|---|---|---|---|---|---|---|
| 1 | Mainly non-porphyritic olivine basalts | The Storr | A | 24 | 23 | Flow 7 thin and weathered |
| | | The Quirang | B | 8 | 8 | |
| | | Near Portree | C | 2 | 2 | |
| 2 | Porphyritic and non-porphyritic olivine basalts | Oisgill Bay | D | 8 | 8 | |
| 3 | Irregular alternations of mugearites and basalts | Beinne na Greine | E | 10 | 9 | Flow 9 poorly exposed |
| 4 | Mugearites and trachytes | Ros a Mheallain | F | 9 | 7 | Badly weathered stream section. Flows 2 & 5 not sampled. |
| 5 | Mugearites and basalts | Osdale River | G | ? | 1 | Poorly exposed group. Confused by many dykes. |

The basal two flows of group 1 were sampled near Portree but the number of flows present between these and those at the Storr is not known. The lavas of this group were sampled at both the Quirang and the Storr to see if it is possible to effect a correlation by comparing the patterns of secular variation revealed by the two sections.

The extrusion of lavas was followed by the intrusion of a plutonic complex now exposed in the southern part of the island. The Cuillins and Blaven are composed of gabbro which has the form of a large sheet or laccolith of complex composition. There appear to have been two phases of intrusion of gabbroic

magma between which was an intrusion of peridotite magma (Douglas Weedon, private communication). The earlier gabbro is unbanded and was sampled in Coire na Banachdich (locality H, Figure 1(a)) along with a xenolith of basalt in the gabbro. The later gabbro is beautifully banded and was sampled at Druim Hain (I).

The last phase in the cycle of igneous activity was the intrusion of a large number of basic dykes which were sampled at the Storr, Oisgill Bay, Osdale and Coire na Banachdich.

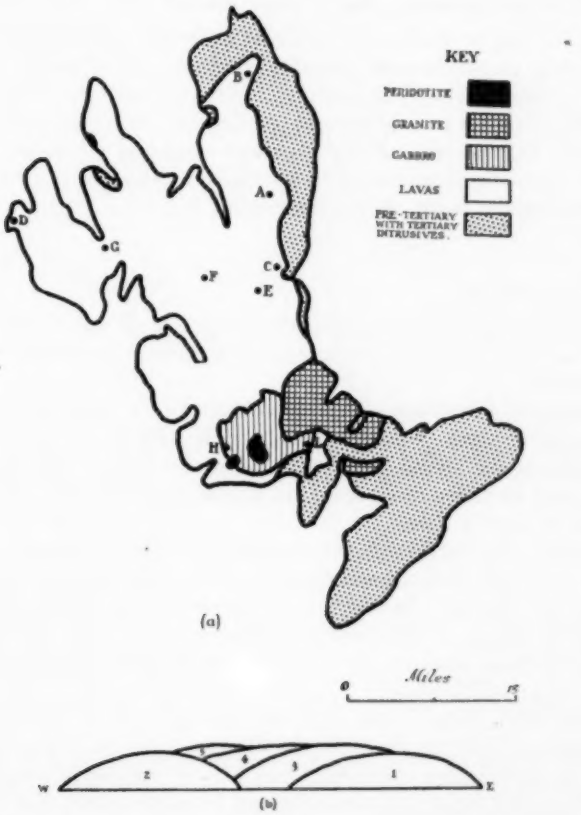


FIG. 1

(a) Geological sketch-map of Skye showing sampling localities A, B, C, ... etc., referred to in the text. (Reproduced from Sheet 1 of the 1/625 000 geological map of Great Britain by kind permission of the Controller, H.M. Stationery Office).
(b) Possible relationship of Skye lava groups.

(b) *Ardnamurchan.*—This area has been through a cycle of igneous activity similar to that of Skye but its plutonic history has been rather more complex. The plutonic rocks form a central intrusion ring complex, now in a fine state of preservation. The ring of banded hypersthene gabbro has been studied in detail by M. K. Wells (1953) and samples were taken from the following three localities described by him: (i) near Sanna Point; (ii) just south of the summit of Beinne na Seilg; and (iii) near Ardnamurchan Point.

(c) *Rhum.*—In the Hallival–Askival area of Rhum some 2 600 ft of rhythmically layered gabbro are exposed (Brown 1957). Each layer consists of a peridotitic rock which passes gradually upwards into an alluvialitic rock. Fifteen major rhythms varying in thickness (average 200 ft) have been recognized and the more resistant, well-exposed, alluvialitic parts of the top eight layers have been sampled.

## 3. Specimen preparation and measurement

In the laboratory the samples were set as *in situ* in plaster of Paris and up to six cylindrical cores of 1·9 cm diameter removed with a diamond impregnated bit fitted on to a standard drilling machine. The cores were then sliced into cylinders 1·9 cm long. These oriented cylinders were wedged into a cubical perspex box and the direction and intensity of the remanent magnetization measured on the Birmingham "spinning magnetometer" (Griffiths 1955). The error in direction of a single measurement made on the instrument is $\pm 1°$.

## 4. Stability criteria

In this work several specimens from each geological unit* have been subjected to one or more of the following tests to decide whether or not the magnetization of the unit is stable.

(a) *Statistical distribution of remanence directions.*—The directions of remanent magnetization of several specimens from each unit have been measured. The magnetization is considered to be stable if the directions of the specimens show a close grouping around a mean direction which is significantly different from that of the present Earth's field.

(b) *Change of direction of magnetization on storage.*—Most specimens were stored for six weeks after measurement with their directions of magnetization at right angles to the Earth's field in the laboratory and then remeasured. Specimens showing little change of direction (less than 3°) were regarded as stable, while those showing large changes were regarded as unstable (Thellier 1951, Blundell 1957).

(c) *NRM/TRM Ratio.*—The ratio of the intensity of the natural remanent magnetization (NRM) to that of the thermoremanent magnetization (TRM) produced by cooling from above the Curie Point in the Earth's field in the laboratory was estimated for at least one specimen from each unit. Specimens with the ratio less than but close to unity were regarded as stable; those for which the ratio is small (less than 0·2) were regarded as unstable; those with the ratio greater than unity are believed to have been affected by lightning (see below).

(d) *Thermal demagnetization.* The intensity of the NRM of rocks decreases with increase in temperature in a manner depending on the stability of the magnetization. Thus, unstably magnetized rocks show a rapid decrease of intensity as the temperature is increased to 50°C or 100°C; while the stably magnetized specimens show the main intensity drop at temperatures much nearer the Curie Point. A small number of thermal experiments was carried out using a non-magnetic furnace designed by Leng at Imperial College, London (Leng 1955). Curves of

* A geological unit is here taken to mean a body of igneous rock which has cooled from above its Curie Point to the air temperature in a time short compared with the period of secular variation (500–1000) years (Hospers 1953). Thus, a lava flow, a group of similar dykes at a site, or a small volume within a large intrusion is treated as a geological unit.

intensity against temperature were obtained during the heating and cooling of the specimen. The heating and cooling curves for both stable and unstable types of specimen are shown in Figure 2(a) and (b) (Nagata 1953).

(e) *A.C. demagnetization.*—The NRM of rocks is generally regarded as consisting of a "hard" (TRM) component of high coercivity, and a "soft" (isothermal remanent magnetization or IRM) component of low coercivity acquired through exposure for long periods of time in the Earth's magnetic field. An indication of stability can be obtained by noting the decrease in intensity of magnetization obtained by subjecting specimens to an alternating field which is smoothly decreased

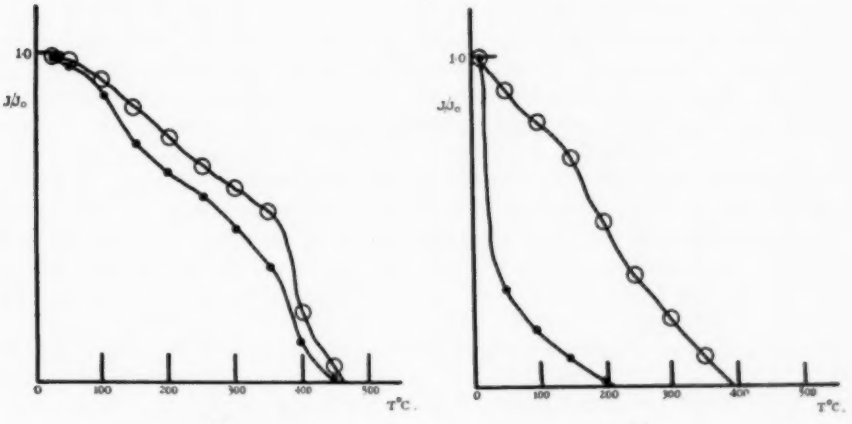


FIG. 2

Ordinate: normalized intensity of magnetization.
Abscissa: temperature (°C)

● heating; ○ cooling; J intensity of magnetization; $J_0$ intensity of magnetization at room temperature.

(a) Thermal demagnetization curves for a stable specimen.
(b) Thermal demagnetization curves for an unstable specimen.

to zero. The form of the curve of intensity of magnetization against the maximum applied alternating field can give a useful indication of stability (Uyeda 1958). Some workers (Creer 1958, As & others 1958) have successfully used alternating fields to remove unstable components from partially stable rocks. An attempt was made in this work to develop an apparatus to achieve such results.

To avoid the development of anhysteretic magnetization (Rimbert 1958) the experiments were carried out in a region in which the Earth's magnetic field was compensated by three Helmholtz pairs mutually at right angles. To allow for inadequate compensation and for any asymmetry in the wave form of the applied alternating field, the specimens were placed in a specially designed holder during demagnetization. The holder was rotated about two mutually perpendicular axes simultaneously while the field was reduced smoothly to zero using a liquid resistance. This arrangement works well for typical rock specimens to which a "hard" component and a "soft" component have been imparted in the laboratory. This is illustrated in Figure 3. The "hard" component was produced by cooling the specimen in the Earth's field from above the Curie Point to room temperature.

The direction and intensity of this magnetization is represented by the points numbered 1. The specimen was then given a "soft" component by placing it in a direct field of 100 oersteds (points 2). The direction and intensity after each application of alternating field are shown by points 3 to 7. It is seen that the directions move along a great circle to that of the TRM when a peak alternating field of 100

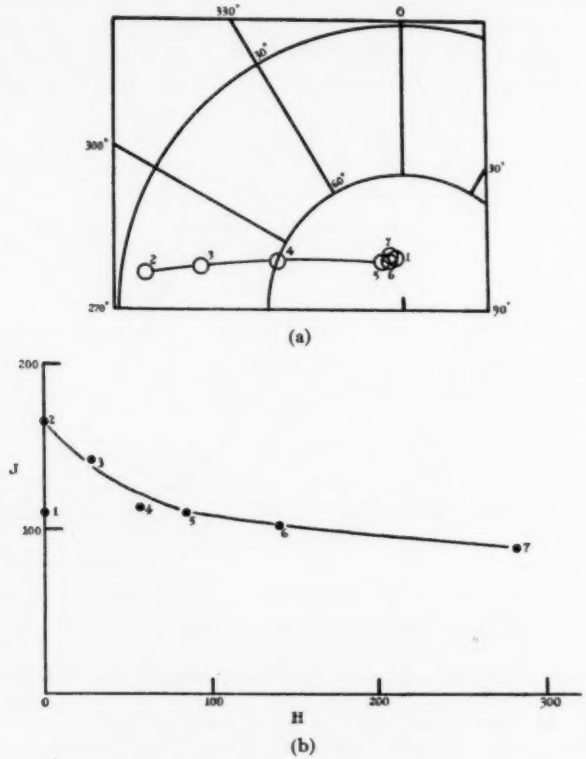


FIG. 3

A.c. demagnetization of an artificial two-component system. 1, thermoremanent magnetization produced in the laboratory; 2, thermoremanent magnetization plus an isothermal component produced by a direct field of 100 oersteds; 3, 4, 5, 6, 7 after application of an alternating field.

(a) Stereographic projection showing directions.
(b) Ordinate: intensity of magnetization (arbitrary units).
Abscissa: maximum applied alternating field (peak values in oersteds).
Variation of intensity with applied alternating field.

oersteds is reached and that the intensity of the TRM is also reached at this field. At higher fields, no change in direction is observed while the intensity of magnetization decreased steadily.

There is little doubt that in the above experiment the "soft" component of magnetization has been successfully removed. However, no specimen with an NRM behaved in this fashion. Figure 4 shows the directions of magnetization of 4 specimens from a single typical sample, which was thought to be partially stable, after successive applications of alternating fields as above. There is no

systematic change of direction. The intensities of magnetization also show no systematic behaviour and increases of up to 200 per cent as well as decreases have been observed after the application of fields up to 400 oersteds (r.m.s.).

It seems that the NRM of these specimens is much more complex than was envisaged and that its true nature must be the subject of further work.

Using the above stability criteria, it is possible to divide the specimens into three groups as follows:

(a) A stable group in which the magnetization directions from a unit are closely grouped around a mean direction nearly opposite that of the present Earth's field (Figure 5) and do not change appreciably on storage. The NRM/TRM ratio is on the average 0·7 to 0·8, and the thermal demagnetization curves are of the stable type. They show little or no change of direction and some loss of intensity when subjected to an alternating field of 400 oersteds (r.m.s.). There is little doubt that the NRM of these specimens has been acquired by cooling through their Curie Points in the Earth's magnetic field. Their magnetization directions thus reflect that of the Earth's field acting when the magnetization was acquired.

(b) An unstable group in which the magnetization directions of the specimens from a unit are widely dispersed (Figure 6) and change by up to 90° on storage. The NRM/TRM ratios are generally small (down to 0·01), and the thermal demagnetization curves are of the unstable type. The magnetization is greatly affected by quite low alternating fields, sometimes a decrease of the order of 50 per cent in intensity is produced by a field of 10 oersteds. The magnetization of such specimens are clearly unreliable and they are therefore of no use for palaeomagnetic purposes.

(c) In the third group, the specimens from within a sample (roughly a 4 in. to 6 in. cube) usually show a close grouping of directions which are not altered on storage. The mean direction of samples from within a unit are, however, quite different (Figure 7). The NRM/TRM ratios are usually greater than unity, values ranging from 1·2 to 35·0 having been observed, and the thermal demagnetization curves are of the stable type. Alternating fields of up to 400 oersteds have little effect on the magnetization. All samples showing these properties have been collected from the tops of hills. They have evidently been magnetized in fields other than that of the Earth, and it seems likely that the strong alternating and direct fields associated with lightning discharges have led to the development of anhysteretic remanent magnetization in these samples. Such samples are obviously of no use for palaeomagnetic purposes and have therefore been neglected.

## 5. Results

The results of the measurements made are summarized in Table 3 in which the total number of specimens measured and the numbers rejected because of instability and lightning effects are given. The mean intensities of magnetization in e.m.u./cc for the stable, unstable and lightning-affected specimens are given for each locality. The mean declinations, inclinations, angles of confidence ($P = 33\frac{1}{3}$ per cent), and dispersion parameters were estimated weighting each unit equally.

The declination and inclinations of individual flows from the Storr and the Quirang are shown in Figure 8, the errors being the values of the semi-angle of the cone of $66\frac{2}{3}$ per cent confidence which gives the spherical equivalent of the standard deviation.

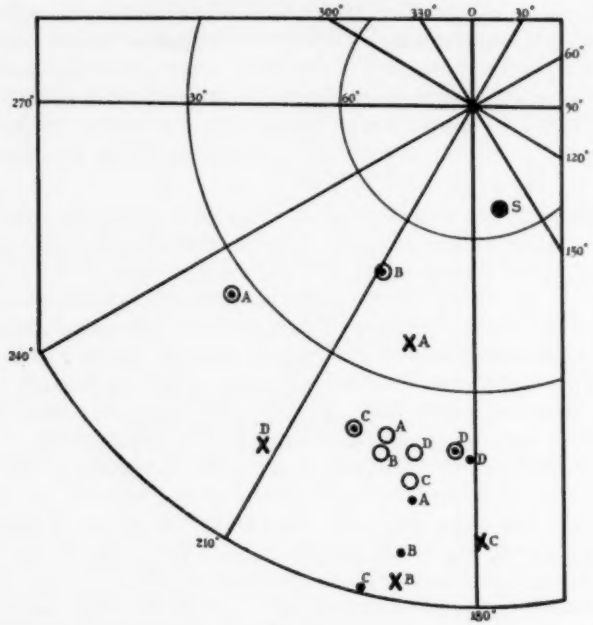


FIG. 4

Stereographic projection showing magnetization directions of specimens A, B, C, and D from the same sample after a.c. demagnetization. ⊙ directions of N.R.M.; ○ after 50 oersteds (r.m.s.); • after 100 oersteds; × after 200 oersteds ● S present south pole (upper hemisphere, all other points are in the lower hemisphere).

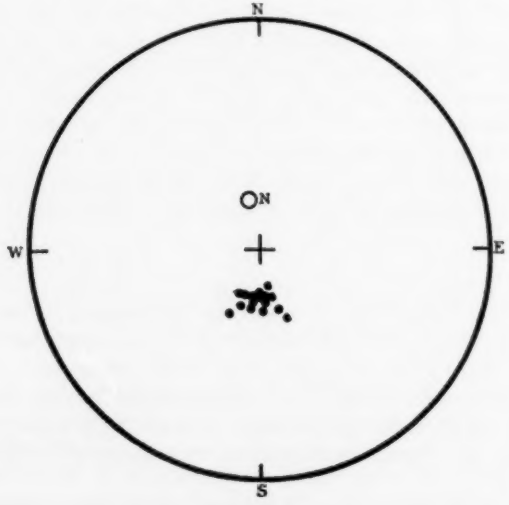


FIG. 5

Stereographic projection showing magnetization directions of specimens from a stably magnetized geological unit. ○N present North pole; ● points in upper hemisphere.

**Table 3**

*Summary of results*

| Rocks | Total Number of specimens | Number of unstable specimens | Number of lightning-affected specimens | Mean intensity of stable specimens e.m.u./cc ($\times 10^{-3}$) | Mean intensity of unstable specimens e.m.u./cc ($\times 10^{-3}$) | Mean intensity of lightning-affected specimens e.m.u./cc ($\times 10^{-3}$) | Number of reliable geological units | Mean declination (degrees from true North) | Mean inclination (degrees from horizontal) | Semi-angle of cone of $66\frac{2}{3}$ per cent confidence | Dispersion parameter |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Storr lavas (no dip correction) | 230 | 11 | 16 | 4·01 | 1·40 | 46·25 | 21 | 191·0 | 61·6 | 2·9 | 44·5 |
| Storr lavas (after dip correction) | 230 | 11 | 16 | 4·01 | 1·40 | 46·25 | 21 | 162·0 | 62·0 | 2·9 | 44·5 |
| Quirang lavas | 75 | 4 | — | 4·02 | 0·33 | — | 8 | 183·0 | 57·3 | 4·5 | 57·3 |
| Oisgill Bay lavas | 103 | 39 | 6 | 5·75 | 0·80 | 1·49 | 7 | 190·3 | 62·3 | 7·1 | 23·2 |
| Beinne na Greine lavas | 106 | 42 | 24 | 3·79 | 1·52 | 50·85 | 6 | 168·2 | 54·0 | 8·8 | 17·9 |
| Bracadale lavas | 83 | 55 | 9 | 1·74 | 2·02 | 7·50 | 3 | 187·8 | 52·4 | 6·2 | 84·8 |
| Osdale lavas | 13 | — | — | 3·54 | — | — | 1 | 166·6 | 61·5 | — | — |
| Portree lavas | 14 | — | 5 | 2·21 | — | 19·97 | 1 | 197·6 | 67·6 | — | — |
| Coire na Banachdich gabbro | 24 | — | — | 4·55 | — | — | 1 | 199·1 | 65·8 | — | — |
| Coire na Banachdich basalt xenolith | 5 | — | — | 2·58 | — | — | 1 | 181·9 | 73·6 | — | — |
| Storr dykes | 22 | 14 | — | 2·13 | 0·48 | — | 1 | 183·0 | 41·0 | — | — |
| Oisgill dykes | 13 | — | — | 2·54 | — | — | 1 | 187·7 | 44·0 | — | — |
| Coire na Banachdich dykes | 4 | — | — | 3·24 | — | — | 1 | 212·1 | 63·4 | — | — |
| Druim Hain gabbro | 11 | — | 11 | — | — | 2·97 | — | — | — | — | — |
| Osdale dykes | 21 | — | — | 1·02 | — | — | 1 | 181·9 | 56·0 | — | — |
| All Skye | 724 | 165 | 71 | 3·81 | 1·39 | 30·55 | 53 | 186·1 | 59·7 | 2·0 | 59·7 |
| Ardnamurchan hypersthene gabbro | 65 | — | 24 | 4·08 | — | 29·61 | 3 | 184·1 | 54·5 | 7·2 | 61·9 |
| Rhum layered gabbro | 38 | — | 11 | 1·22 | — | 136·33 | 4 | 181·0 | 55·1 | 4·6 | 97·1 |

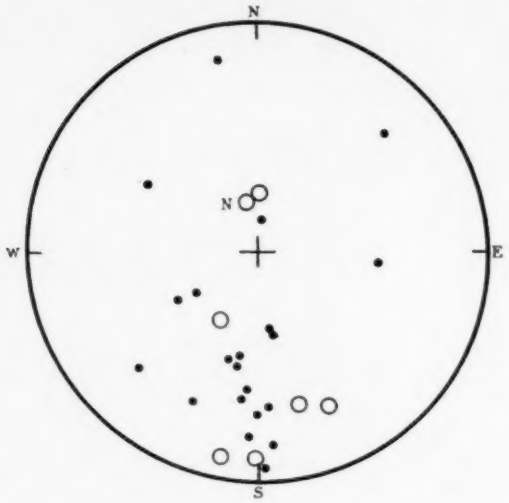


FIG. 6.—Stereographic projection showing magnetization directions of specimens from an unstably magnetized geological unit. ○N present North Pole; • points in upper hemisphere; ○ points in lower hemisphere.

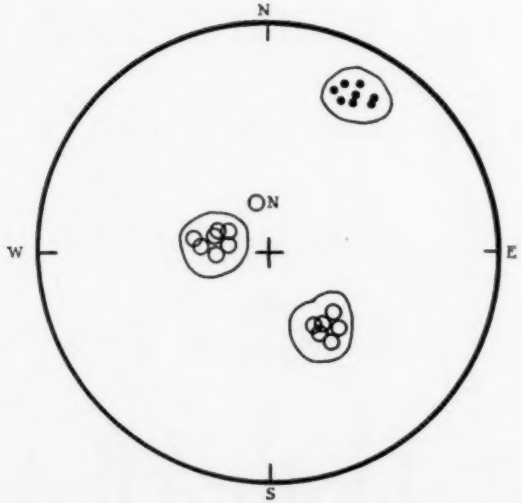


FIG. 7.—Stereographic projection showing magnetization directions of specimens from three samples of a unit affected by lightning. Specimens from the same sample are enclosed by closed curves. ○N present North Pole; ○ points in lower hemisphere; • points in upper hemisphere.

The mean direction of magnetization of each of the Skye units is shown in Figure 9.

The ancient pole positions estimated from Skye, Ardnamurchan and Rhum in present-day geographical coordinates are given in Table 4.

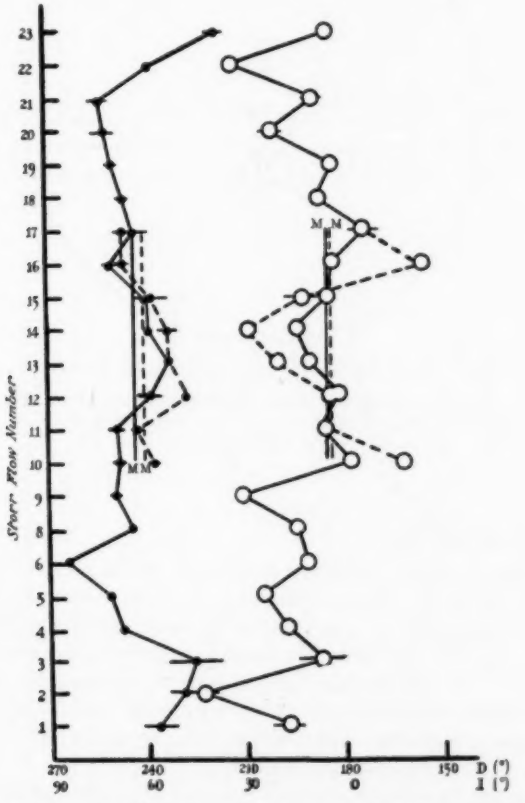


FIG. 8. Secular variation from the lavas of the Storr and Quirang. ● inclination; ○ declination; ——Storr; - - - - Quirang; M mean.

## Table 4

*Ancient pole positions*

| Location | Lat. | Ancient Pole Position Long. | *dp* | *dm* |
|---|---|---|---|---|
| Skye | 70°·8 N | 159°·6 E | 3°·8 | 5°·1 |
| Ardnamurchan | 68°·1 N | 165°·0 E | 15°·4 | 21°·9 |
| Rhum | 68°·5 N | 150°·9 E | 14°·0 | 13°·4 |

The half angles of the oval cones of confidence are here estimated at the 95 per cent level for easy comparison with the results of earlier workers given in Table 1.

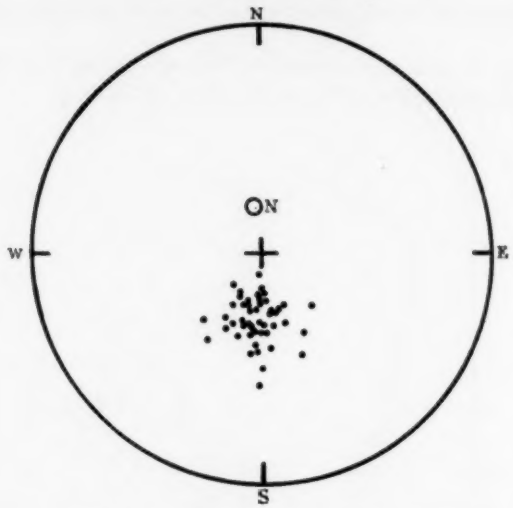


FIG. 9. Stereographic projection showing magnetization directions of each of the Skye geological units. ○N present North Pole; • points in upper hemisphere.

## 6. Discussion of results

It is seen from Table 3 that about 10 per cent of the Skye rocks (all from hill-tops) have been affected by lightning and about 20 per cent possess unstable magnetization. In some instances, e.g. most of the specimens from the stream section near Bracadale, the rocks possessing unstable magnetization are more weathered than the average. It is also evident that, in general, the unstable rocks have lower intensities of magnetization than the stable ones, while those affected by lightning have much higher intensities. None of the gabbros was found to be unstable but about 40 per cent were affected by lightning.

The lavas at the Storr have a westerly dip of 14° (determined with the aid of orientated photographs). The base of the lavas all round the island is not far from sea level and Dr F. W. Anderson of the Geological Survey has estimated that the lavas have not been tilted by more than $\frac{1}{4}$° since they were formed. This is supported by the fact that the lavas of Oisgill Bay, which are horizontal, and are probably of the same age, have the same mean magnetization direction as the Storr lavas provided they are not corrected, but the difference is large if a correction is made. It is, therefore, concluded that no dip corrections are necessary for the Skye rocks.

The declinations and inclinations of the individual flows of the Storr and the Quirang are shown in Figure 8. The differences in magnetization direction between the flows are attributed to secular variation. The maximum departure is seen to be within 30° of the mean in declination and 25° in inclination, values which are compatible with the views of other workers. An attempt was made to correlate the lavas of the two sections by comparing their secular variation patterns by eye; the best fit was obtained by equating the basal Quirang flow with the tenth flow of the Storr. In addition, for each possible correlation (assuming that all the

Quirang flows are represented at the Storr) two correlation coefficients were estimated—one for the eight pairs of declination values, and one for the eight pairs of inclination values. The fifteen pairs of coefficients, obtained by equating the basal Quirang flow with each of the lower fifteen Storr flows in turn, are plotted in Figure 10 as ordinates against the Storr flow number with which the basal Quirang

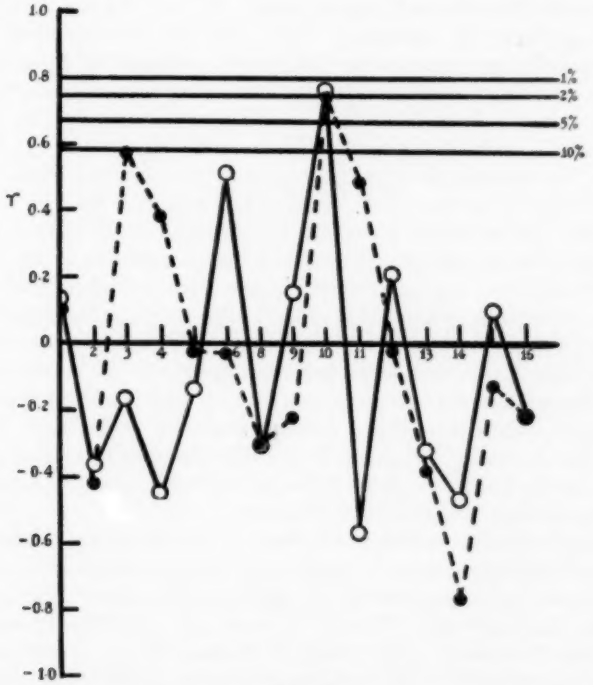


FIG. 10

Ordinate: coefficient of correlation
Abscissa: Storr flow number with which the basal Quirang flow is correlated.

Correlation of lavas of flows of the Storr and the Quirang. ——inclinations; - - - - declinations. The 1, 2, 5 and 10 per cent levels of significance are shown.

flow is correlated. For both parameters the maximum value of the coefficient is obtained when the basal Quirang is correlated with the tenth flow of the Storr, and it is seen that in both there is only a 2 per cent probability that this is due to chance. It is also noticeable that this is the only postulated correlation for which the correlation coefficient is of a value usually regarded as significant (Paradine & Rivett 1953). The mean declination and inclination of the two equivalent sets of flow are seen to be almost identical. The above is quite strong evidence for concluding that if there is complete overlap of the two sets of flows, the basal Quirang flow must be equated with the tenth Storr flow.

It is not possible to effect a similar correlation between the lavas of the Storr and Oisgill Bay as they belong to different groups which probably do not contain flows which were formed exactly contemporaneously.

The differences between the mean magnetization directions for the various sites given in Table 3 are attributed to incomplete averaging of the secular variation. The similarity of the magnetization directions shown by the dykes of the Storr, Oisgill Bay, and Osdale probably means that they were formed at the same time, which is different from that at which the Coire na Banachdich dykes were formed.

A striking fact about the directions of magnetization is that all the stably magnetized rocks show reverse magnetization. As none of the rocks shows self-reversing properties in the laboratory, the reverse magnetization is believed to be due to the rocks being formed when the Earth's magnetic field was reversed. It is difficult to estimate the time covered by this. Hawkes (1916) estimated that the average time between flows in Iceland is about 1 000 years, but Hospers (1954) argues that this is probably an underestimate, and suggests a time nearer 10 000 years. Since, for example, in Skye there are over 50 flows, the Earth's field must have been reversed for a time of the order of half a million years.

It is highly probable that by taking the mean of the fifty-three Skye units examined, the effect of secular variation will be averaged out. The ancient pole position estimated from this mean therefore corresponds to that of the geographic pole position at the time of formation of the rocks, i.e. probably Lower Miocene. This position (Table 4) is significantly different from that of the present geographic pole. It therefore seems that there has been some relative movement between North-west Europe and the pole since this time. There is still a great need, however, for further results from other Tertiary areas (e.g. Greenland, Spitzbergen, Jan Mayen, The Faroes). This, along with further data from America and accurate dating (e.g. by the Argon method) would provide important information about relative movement between the two continents.

The gabbros from Rhum and Ardnamurchan (excluding the lightning-affected samples) were found to be reversely magnetized, but not enough sites were sampled to give an accurate estimate of the pole position in either case. It is interesting to note, however, that they were found to be eminently suitable for palaeomagnetic purposes (Blundell & Read 1958, Gough & Niekerk 1959). There is scope for much more work on gabbros, for by suitable sampling and thermal studies in the laboratory, they can provide a more continuous pattern of secular variation than can a succession of lava flows (Jaeger 1957). The pole positions estimated (Table 4) from Rhum and Ardnamurchan are quite close to that estimated from Skye, and the fact that they are also reversely magnetized and that no normally magnetized rocks were found, could mean that the rocks studied from all three areas were formed within the same period of reversal of the field, i.e. within the same half million years.

## 7. Acknowledgments

The writer is grateful to Professor J. McG. Bruckshaw and Dr D. H. Griffiths for suggesting this work, which has been carried out with the constant help of Dr R. F. King and Dr Griffiths.

He is also grateful to many others whose help made this work possible, especially Dr F. W. Anderson and Dr J. B. Simpson of the Geological Survey of Great Britain and Dr Douglas Weedon of the University of Glasgow for access to unpublished geological works; Dr J. A. Clegg and Dr C. W. F. Everitt of the Physics Department, Imperial College of Science, London for generous use of and kind assistance with the non-magnetic furnace; Dr A. I. Rees whose help in the field

has been invaluable; Professor F. W. Shotton and other members of the Geology Department of the University of Birmingham, for generous use of departmental facilities; and Mr S. Bright for invaluable technical assistance.

Finally, he wishes to thank the Trinidad Petroleum Development Company for an award during the first half of the investigation and to the University of Birmingham for a grant during the second half.

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# The use of Rapid Digital Computing Methods for Direct Gravity Interpretation of Sedimentary Basins

M. H. P. Bott

(Received 1959 September 23)

*Summary*

A rapid digital computing method for the direct calculation of the shapes of two-dimensional sedimentary basins of known density contrast from their residual gravity anomalies is described. In conclusion an example of the use of the method in determining the shape of the Dumfries New Red Sandstone basin (South Scotland) is given.

## 1. Introduction

Two rapid computing methods for the interpretation of effectively two-dimensional gravity anomalies have recently been developed. The first method enables the calculation of the theoretical anomalies of specified two-dimensional shapes bounded by a connected polygonal series of straight lines. A programme similar in principle has already been developed and described by Talwani & others (1959) and therefore this method need not be described.

The second method enables the shapes of two-dimensional sedimentary basins to be obtained directly from the regionally corrected gravity anomalies profile. This is logically a development from the first method and is based on the principle of successive approximation for the removal of residuals. This method is described in the following account and in conclusion an example of its use is given.

## 2. Calculation of shapes of sedimentary basins

In general an observed gravity profile can be caused by an infinite number of different mass distributions. The consequent uncertainty in quantitative interpretation can sometimes, however, be reduced or even eliminated if certain assumptions may legitimately be made from our geological knowledge. One such set of circumstances whereby a unique interpretation is possible arises when a gravity anomaly can be entirely attributed to a sedimentary basin for which the density of the sediments in relation to the surrounding and underlying rocks is known. In practice it is possible to calculate the shape of the floor of such an effectively two-dimensional sedimentary basin from the regionally corrected gravity anomalies by a process of trial and error which can be readily adapted for rapid digital computing methods.

The method which has been developed depends on the division of the total width of the outcropping sedimentary basin into a series of two-dimensional strips which need not necessarily be of uniform width (the half width of the $i$th strip will be represented by $w_i$). The regionally corrected gravity anomaly observed at the centre point of each strip is obtained from the profile and will be denoted by $(A_{obs})_i$. The object is then to obtain the pattern of sediment thicknesses beneath

the strips (in the form of two-dimensional rectangular blocks) which can account for all the values of $A_{obs}$.

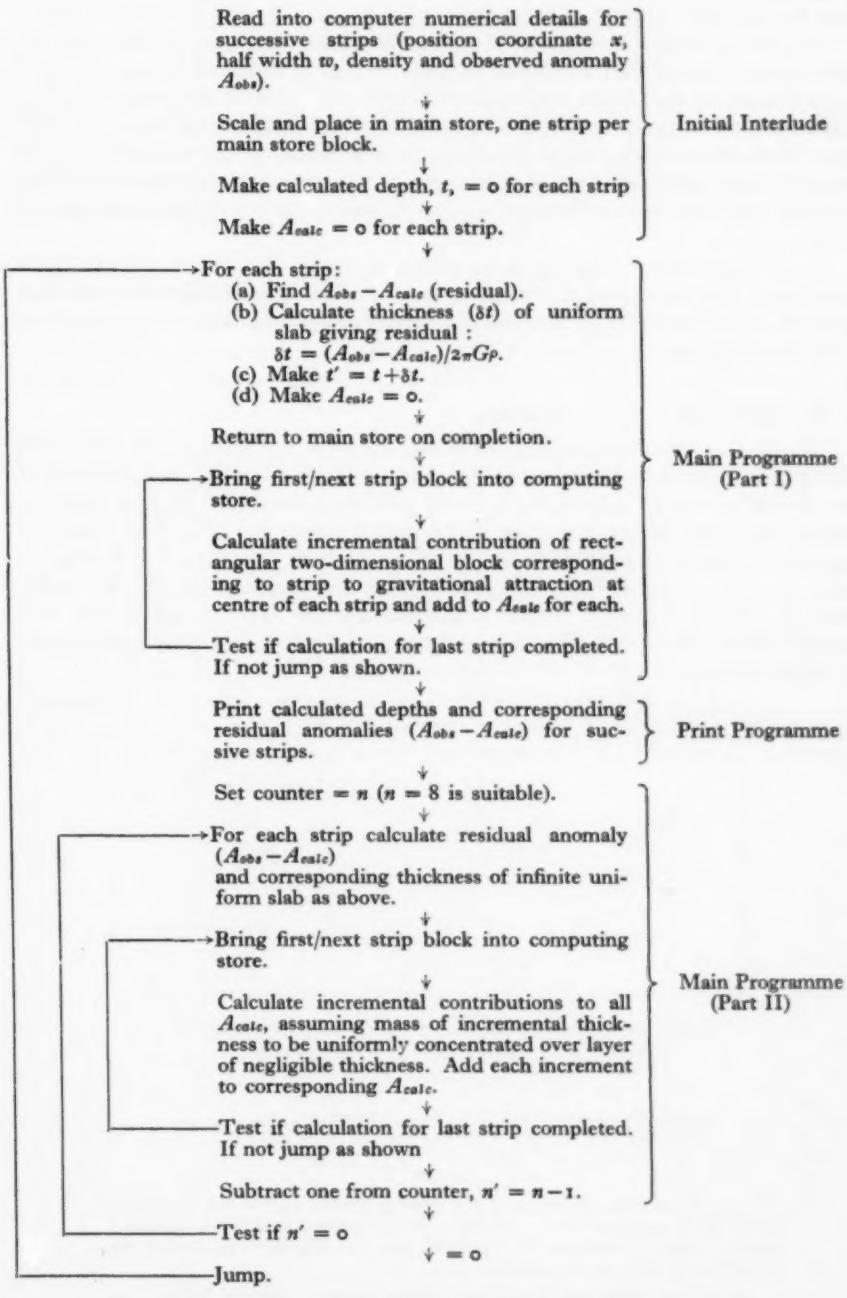
The first approximation to the shape is obtained by the computer by assuming the thickness of sediments beneath each strip to be given by the thickness of a horizontal infinite slab ($t_i$) of the assumed density contrast needed to give $(A_{obs})_i$. The calculated values of gravity for this model $(A_{calc})_i$ are obtained by considering the sediments beneath each strip to be in the form of a two-dimensional rectangular block, depth $t_i$ and width $2w_i$, and by summing the contributions of all the blocks for each position. Each individual calculation is based on the formula given by Heiland (1940, p. 152) for the gravitational attraction of a vertically sided two-dimensional block. The residual anomalies are obtained by subtracting the calculated values of the anomalies for the model from the observed values. To this stage an approximate model has been constructed, the exact anomalies over the model have been calculated and the residual anomalies, indicating the degree of misfit of the model, have been obtained by subtracting calculated from observed anomalies.

The succeeding stages of the calculation aim at progressive reduction and removal of the residuals by repeated alteration of $t_i$ and recalculation of residuals. Each stage of the adjustment begins with the estimation of the alteration of the depth of sediments beneath each strip ($t_i$) by calculating the thickness of an infinite slab of the assumed density contrast needed to account for the corresponding residuals. Having approximately adjusted the shape in this way, the anomalies can be recalculated in either of two ways, both of which are used in the programme. The first method involves a complete recalculation of the anomalies for the new model as has been outlined above. The second method depends on the direct adjustment of the previous calculated values for the gravitational effect of the incremental thickness of sediments added or removed from the bases of the rectangular blocks: since an approximation can be conveniently used in this second method it is much faster.

A "flow diagram" for the programme, as prepared for the Ferranti Pegasus computer, is shown below ($G$ representing the gravitational constant). The following paragraphs give a short explanation of the divisions of the programme.

The purpose of the *initial interlude* is to read into the computer the numerical information, suitably scaled and arranged. The main programme, following the initial interlude, is divided into two parts. *Part I* provides for an adjustment of the depths of the rectangular blocks in accordance with the residuals, followed by a full and rigorous recalculation of the gravitational effect of the adjusted model. The residuals are then recalculated. If desired the calculated depths and the corresponding residuals may be printed at this stage. *Part I* is used both at the beginning of the calculation when the approximation involved in Part II would introduce too great errors and also before printing the results.

*Part II* of the main programme similarly begins with the alteration of the depths of the rectangular blocks in accordance with their residuals. A complete recalculation of the theoretical anomalies is, however, avoided by calculating directly the gravitational effect of the increments and by adjusting the previous values by these amounts. An approximation to the effect of the incremental rectangular block can be made by assuming the mass to be concentrated uniformly over a two-dimensional plate. The errors are negligible under normal circumstances and in consequence Part II becomes four times faster than Part I. It has been


Read into computer numerical details for successive strips (position coordinate $x$, half width $w$, density and observed anomaly $A_{obs}$).
Scale and place in main store, one strip per main store block.
Initial Interlude
Make calculated depth, $t$, = 0 for each strip
Make $A_{calc}$ = 0 for each strip.
For each strip:
(a) Find $A_{obs} - A_{calc}$ (residual).
(b) Calculate thickness ($\delta t$) of uniform slab giving residual : $\delta t = (A_{obs} - A_{calc})/2\pi G\rho$.
(c) Make $t' = t + \delta t$.
(d) Make $A_{calc}$ = 0.
Return to main store on completion.
Bring first/next strip block into computing store.
Calculate incremental contribution of rectangular two-dimensional block corresponding to strip to gravitational attraction at centre of each strip and add to $A_{calc}$ for each.
Test if calculation for last strip completed. If not jump as shown.
Main Programme (Part I)
Print calculated depths and corresponding residual anomalies ($A_{obs} - A_{calc}$) for successive strips.
Print Programme
Set counter = $n$ ($n$ = 8 is suitable).
For each strip calculate residual anomaly ($A_{obs} - A_{calc}$) and corresponding thickness of infinite uniform slab as above.
Bring first/next strip block into computing store.
Calculate incremental contributions to all $A_{calc}$, assuming mass of incremental thickness to be uniformly concentrated over layer of negligible thickness. Add each increment to corresponding $A_{calc}$.
Test if calculation for last strip completed. If not jump as shown
Subtract one from counter, $n' = n - 1$.
Test if $n'$ = 0
= 0
Jump.
Main Programme (Part II)

Flow diagram for the programme.

found convenient to repeat Part II eight times before re-entering Part I in preparation for printing.

In practice it has been found that usually about ten stages of successive approximation (one of Part I followed by eight of Part II followed by one of Part I) are sufficient to reduce the residuals well below the level of observational error. Less rapid convergence is usually an indication of an observational error or a local disturbing effect. It has been found that a calculation of ten stages for a sedimentary basin subdivided into ten strips takes about three minutes on the Ferranti Pegasus computer used. The time is proportional to the square of the number of strips.

The programme, including subroutines, occupies about forty blocks (each containing eight words) of the main store. In addition each strip uses one main store block. Since the Pegasus main store contains 640 blocks there is no problem over storage space.

## 3. An application of the method

The method has been used to obtain the shapes of the floors of two New Red Sandstone basins near Dumfries in the south of Scotland. The interpretation of the gravity survey of this region is being published elsewhere (Bott & Masson-Smith 1960) but the interpretation of the profile across the Dumfries trough is reproduced here as an example of the use of the method. The New Red Sandstone rocks are about 0·4 g/cm$^3$ lower in density than the surrounding Silurian rocks. Both the observed gravity profile (corrected for the regional effect) and two interpretations, based on different assumptions of density distribution, are shown in Figure 1 below.

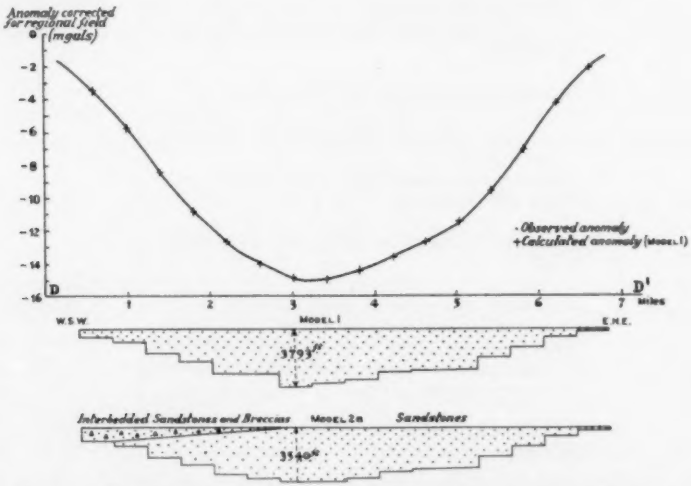



FIG. 1.—The interpretation of the gravity profile across the Dumfries New Red Sandstone basin. Model 1 has been calculated for an assumed uniform density contrast of 0·4 g/cm$^3$ and the calculated anomalies are shown as crosses on the profile.

Model 2 takes into account certain denser breccias which are assumed to be distributed as shown and to have a density of 0·2 g/cm$^3$ lower than the Silurian rocks. The residuals are negligible apart from one value of 0·3 mgal where convergence is poor.

Fuller details of the interpretation are given in the above-mentioned paper, where interpretations with varying strip widths are also shown.

## 4. Possible extensions of the method

It is thought that the method could fairly readily be adapted to basins of finite length by the use of end-effect corrections. A disadvantage is that the final shape needs to be expressed in terms of rectangular blocks but it is feasible that a similar programme could be developed for a polygonal outline to the floor.

## Acknowledgments

The programmes described have been developed at the Durham University Computing Laboratory and grateful acknowledgment is made to the staff for assistance in locating programme errors during development.

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# An Equation of State for the Core of the Earth

**Leon Knopoff and Gordon J. F. MacDonald**

(Received 1959 October 5)

*Summary*

Recent shock wave measurements upon the compressibility of iron and eight other metals, at pressures up to five megabars, permit an investigation of the equation of state of the Earth's core. The density of iron at $T = 0$ at 1·4 megabars (core–mantle boundary pressure) is 11·8. The density at the core boundary is estimated to be between 9·1 and 10·1, depending upon the particular Earth model. The temperature correction is small. The discrepancy can only be resolved by stating that the core is not pure iron, but rather that it contains significant amounts of alloying elements of lower atomic number than iron. The seismic velocity in pure iron at core pressures is also significantly different from the velocity in the core and also indicates the existence of lighter components within the core. A material of mean atomic number 23 in the core is consistent with the shock wave velocity and density measurements and with seismic observations.

## 1. Introduction

The discovery of the core of the Earth by Oldham in 1906 has since provoked much speculation as to the physical and chemical properties of the core. The difficulty of determining, with any degree of definiteness, the physical and chemical nature of the Earth's core is connected with the difficulty of compressing surface materials to the pressures of the Earth's core. The pressures and densities in the Earth's core are reasonably well known (Bullen 1953). Bullard (1957) investigated permissible density distributions within the Earth and concluded that the assigned density of the core at the core–mantle boundary (9·7 g/cm$^3$) should not be in error by more than 0·5 g/cm$^3$. If ordinary materials, taken from the surface of the Earth, could be compressed to core pressures and elevated to core temperatures, then the densities could be compared and a selection made from among the likely candidates for the composition of the core. Until recently, the highest pressures obtained in the laboratory for jacketed specimens have been approximately 100 000 atmospheres (Bridgman 1946); these pressures correspond to a depth of only 250 km in the Earth.

It was suggested by Jensen (1938) that the Thomas–Fermi model for the degenerate state of atoms at extremely high pressures might provide an indication as to the state to which ordinary Earth materials would be compressed at these extreme pressures. Jensen suggested the possibility that laboratory compressibilities of Earth materials could be extended to the Thomas–Fermi equations of state as

asymptotes. At that time, little high pressure data on compressibility of Earth materials was available. Elsasser (1951) took much of the data of Bridgman and extrapolated it to the Thomas–Fermi model and was able to assign an atomic number for the core of the Earth; the atomic number he obtained was 29. This number should be compared with 26 for iron and 28 for nickel. Ever since a suggestion by Wiechert, it has been more or less accepted that the Earth's core is principally iron–nickel.

Bullen (1952) criticized Elsasser's extrapolation of the high pressure results to low pressures and showed that an Earth model based on Elsasser's calculations becomes internally inconsistent. Bullen concluded that the atomic number to be associated with the material of the outer portion of the Earth's core should be of the order of about 23, a figure somewhat less than the atomic number for iron.

The extreme gap in pressure between the laboratory data of Bridgman and the region of validity of the high pressure equation of state of the Thomas–Fermi model makes the interpolation between these two regions extremely uncertain. Knopoff & Uffen (1954) suggested that the interpolation can be substantially improved by extending the high pressure data by the use of the Birch–Murnaghan semi-empirical theory of finite strain. This procedure gave an improved interpolation between the experimental data and the high pressure theoretical models. The results obtained showed again that the atomic number associated with the core was less than that of iron, and in substantial agreement with the value of 23 obtained by Bullen.

MacDonald & Knopoff (1958) compared the abundances of the elements in the solar system, meteorites and solar spectra. They showed that a significant amount of silicon could not be accounted for on the assumption of the preservation of the iron plus magnesium to silicon ratio within the solar system. If this excess silicon is assumed to be present in the core of the Earth, then again the atomic number of such an iron–silicon mixture in the core will be of the order of 23, a number in substantial agreement with the previous estimates.

Other models have been suggested for the composition of the core. The most notable of these include the model of Ramsey (1948, 1949, 1950) who suggested that the seismic discontinuity at the core represents a phase transition from a solid phase of olivine to a metallic liquid phase of the same material. A second and alternate hypothesis which challenges the assumption of an iron core was presented by Kuhn & Rittman (1941) who proposed a core made up of undifferentiated solar matter rich in hydrogen. However, it was shown by Wigner & Huntington (1935) and Kronig, de Boer & Korringa (1946) that significant amounts of hydrogen at the core pressures would yield a material whose density was not sufficiently great compared with that obtained in the core of the Earth. Birch (1952) reviewed the latter two hypotheses and concluded that the core is indeed mainly iron–nickel, although he noted that the density of the core is perhaps 10 to 20 per cent less than that of iron or iron–nickel at core conditions.

## 2. High pressure experiments

All speculations as to the composition of the core must inevitably be compared with high pressure experiments. The results of compressibility measurements will ultimately determine which of the available constituent materials for the Earth's core can be compressed to the density appropriate to the core.

Recently, measurements have been made of the compressibility of iron and several other metals to pressures of the order of 4 million atmospheres (Altshuler & others, 1958a,b). These measurements of compressibility, to pressures of an order of magnitude higher than any hitherto reported, provide the comparison sought for. The measurements have been made using the technique of shock waves obtained from the detonation of high explosives in the vicinity of a sample. A summary of the development of the method, the methods of interpretation of the data, and results of measurements to pressures of the order of 400 kb* is given by Rice, McQueen & Walsh (1958).

In order properly to interpret the data, the equation of state as determined from the shock wave data must be reduced to a reference temperature. The temperatures in the shock front are not generally known and some auxiliary measurements must be made. The equation of state in the shock front is the Hugoniot, a condition which interrelates the conservation of momentum and energy in the shock front; this is neither an adiabatic nor an isothermal condition. The pressure–density data obtained in the shock wave experiments can be reduced to those at absolute zero if Grüneisen's ratio $\gamma$ is known. Grüneisen's ratio is

$$\gamma = \frac{\alpha}{c_v \beta \rho} \tag{1}$$

where $c_v$ is the specific heat at constant volume, $\rho$ is the density, $\alpha$ is the coefficient of thermal expansion and $\beta$ is the compressibility. If $P_h(\rho/\rho_0)$ is the Hugoniot pressure at density $\rho$, then the pressure at absolute zero $P_c$ can be written as

$$P_c = -\tfrac{1}{2}\gamma^2(\rho/\rho_0)^{\gamma+1} \int_1^{\rho/\rho_0} P_h(x)(x-h)x^{-\gamma-2}dx - \tfrac{1}{2}\gamma\left(\frac{\rho}{\rho_0} - h\right)P_h(\rho/\rho_0) \tag{2}$$

where $h = (2/\gamma) + 1$ and Grüneisen's ratio is taken to be constant over the pressure range of interest (Altshuler & others). $\rho_0$ is the density at zero pressure. The reduction of the data therefore depends upon a measurement of Grüneisen's ratio. Altshuler & others, by a rather ingenious experiment, have directly measured Grüneisen's ratio at approximately 1 million atmospheres. Grüneisen's ratio for iron at this pressure is about 1·6. The measurement, involving the compressibility of metallic iron in a dense state and metallic iron in a "porous" state, has not been carried out for the other materials subjected to the high shock experiments. Therefore, for the materials, copper, zinc, silver, cadmium, tin, gold, lead and bismuth, also compressed by shock waves in the Hugoniot pressure range of 400 kb to 4 Mb an estimate of Grüneisen's ratio has to be made.

The equation of state for iron reduced to absolute zero (Figure 1) shows that the experimental data yield a density somewhat in excess of the density of the core at the core–mantle boundary pressure. Corrections have to be made to the experimental data to account for (1) the thermal expansion of iron to the temperatures of the core, and (2) the volume change upon melting. It will be shown that these two corrections are insufficient to account for the discrepancy between the experimental determination of the density of iron at the core pressure and the density of the Earth at this same pressure.

* 1 b (bar) = $10^6$ dyn/cm$^2$.

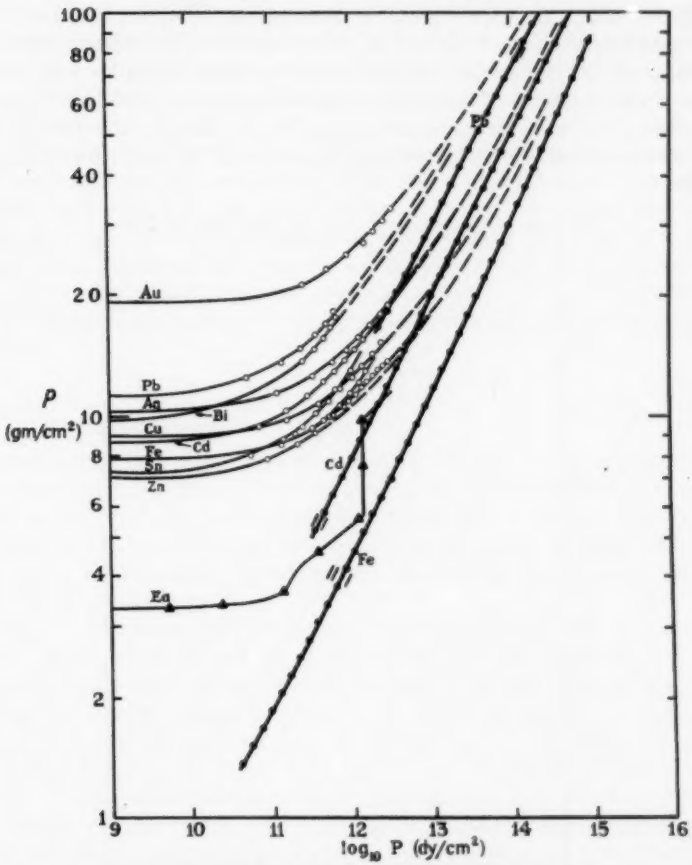


FIG. 1.—Experimentally determined equation of state of nine elements (open circles) reduced to absolute zero, compared with the Thomas–Fermi equation of state at absolute zero (closed circles) and the equation of state for the Earth on Bullen's model (triangles). Extrapolation of the experimental data (dashes) is obtained by integration of the velocity equation of state.

## 3. Grüneisen's ratio at high pressure

The reduction of the Hugoniot equation of state to an equation of state at absolute zero depends very critically upon the value of the Grüneisen ratio $\gamma$. The Grüneisen ratio must be evaluated in the pressure range of interest, and therefore may have little relationship to its value at low pressures. The correction to the experimental data is given in Equation 2. The derivation of Equation 2 is made under the assumption that $\gamma$ is a constant over the pressure range of interest. Unfortunately, $\gamma$ is not a constant over the entire pressure range, but is itself a function of the state of the material (Rice, McQueen & Walsh 1957).

Grüneisen's ratio at pressures of the order of megabars has been measured experimentally by Altshuler & others. No evaluation of this parameter has been made for the other eight materials. In order to determine the value of $\gamma$ for the

materials for which no experimental measurements exist, we use the technique already used in preceding studies of equations of state. We calculate the value of this parameter for extremely high pressures by determining its value from the Thomas–Fermi equation of state. We then attempt to correct these values in the intermediate pressure range using some experimental data; in this case the experimental determination of $\gamma$ for iron is used to correct the value of $\gamma$ for the remaining materials.

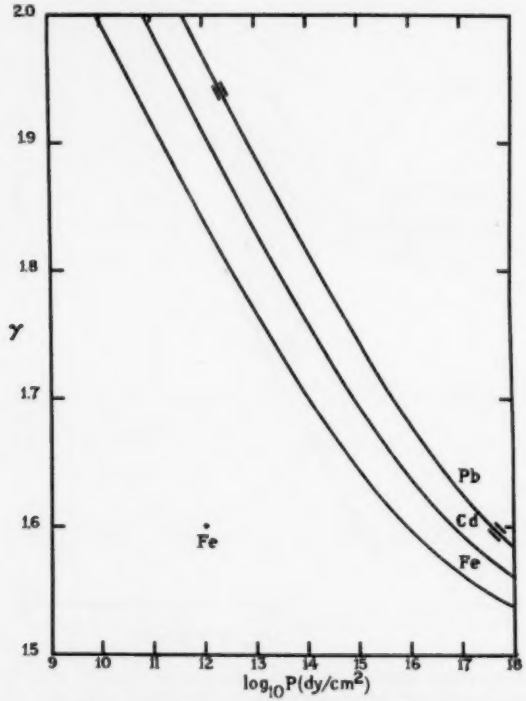


FIG. 2.—The variation of Grüneisen's ratio as a function of pressure for the Thomas–Fermi model of lead, cadmium, and iron. The experimental determination by Altshuler & others is shown as a point.

Although Equation 2 was derived under the assumption that $\gamma$ is a constant, the corrections to the data at low pressures, pressures of the order of 400 kb or less, are not significant and it will suffice to determine a single value for Grüneisen's ratio in the pressure range of the order of 2 or 3 Mb. This should be accurate enough to cover the entire range of the experimental data of Altshuler & others, taken over the pressure range from zero to 4 Mb.

In order to determine the quantum-mechanical values of $\gamma$, we need not only the pressure–density equation of state at absolute zero, but also the way in which the state functions vary with temperature. This is necessary in order to be able to evaluate $c_v$ and $\alpha$. Expansions of the state functions, valid for low temperatures, have been given by Gilvarry (1954).

The quantum-mechanical calculation for $\gamma$ yields numbers, in the pressure range where the quantum-mechanical model becomes significant, of the order of the values of $\gamma$ determined experimentally at low pressures (Figure 2).

It is now not a difficult matter to construct the correction to the quantum-mechanical values for $\gamma$ as a function of pressure, as in Figure 2. From these values we can now select values of $\gamma$ appropriate to the pressure range of the order of megabars and apply the corrections to the Hugoniot equations of state by the method of Equation 2.

The experimental data of Altshuler & others, have been taken for nine materials ranging in atomic number from that of iron (26) to bismuth (83). The metals experimentally measured fall into three groups of atomic numbers: iron (26), copper (29), zinc (30); silver (47), cadmium (48), tin (50); gold (79), lead (82), bismuth (83). The Thomas–Fermi data on all the figures have been plotted only for representative members of each of the three groups; the curves for the remaining elements are sufficiently close to those for which the appropriate curves have been drawn.

## 4. The velocity equation of state

The experimental data of the Hugoniot equation of state for the nine materials have been corrected using Equation 2 and a value of $\gamma$ as determined in the preceding section. The values of $\gamma$ used all fall in the range 1·3 to 1·6. The experimental equations of state at absolute zero have been drawn as the lines with open circles in Figure 1, along with their quantum-mechanical counterparts, valid at extremely high pressures.

It is now possible to interpolate between the experimental data for the nine materials and their quantum-mechanical asymptotes. In order to do this, we construct a velocity equation of state relating velocity in the fluid state of the metal to the pressure. For pressures less than 2 or 3 Mb numerical differentiation has been applied to the corrected experimental data. At extremely high pressures numerical differentiation has been applied to the Thomas–Fermi equation of state. These curves have been drawn in Figure 3. It is now seen that the velocity equation of state determined from the experimental data approaches the quantum-mechanical velocity equations of state at pressures much lower than those at which the corresponding density equations of state approach their quantum-mechanical asymptotes. It would appear that the experimental velocity equation of state approaches the quantum-mechanical equation of state at pressures less than 1 Mb. Thus the quantum-mechanical velocity equation of state can be used at pressures in excess of about 1 Mb. Using the experimentally determined density equation of state, numerical integration can now take place using the quantum-mechanical velocities at the appropriate pressures and the extension of the experimental curves can be made. This has also been done in Figure 1; the dashed curves represent the numerical integration of the experimentally determined density equations of state for the nine materials where the numerical integration has taken place using the quantum-mechanical values of the velocities for each of these materials.

The velocity distribution in the Earth can also be compared with the experimental determinations of the velocities in the nine metals since the pressure range is similar. In Figure 3, the velocity in the Earth has been compared with the velocities at absolute zero for the nine metals. An expansion of the region of interest, pertinent to the core of the Earth, has been made in Figure 4. The velocity distribution in the core of the Earth seems to be in excess of the experimental values of the velocity in iron at absolute zero and is roughly identical to the Thomas–Fermi

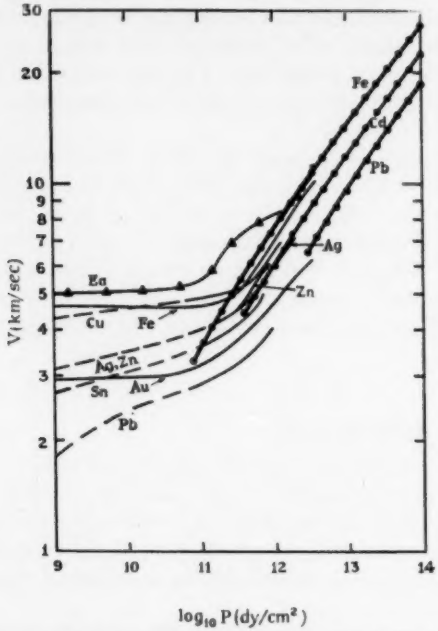


FIG. 3.—The variation with pressure of the velocity of sound in the fluid state. Velocities derived from the Thomas–Fermi model (closed circles) are compared with those obtained from the reduced data of shock wave measurements (solid curves). The seismic velocity distribution given by Bullen (triangles) is also shown for comparison. The dotted curves represent the extrapolation to the zero pressure value of $V = (\alpha^2 - \frac{4}{3}\beta^2)^{\frac{1}{2}}$.

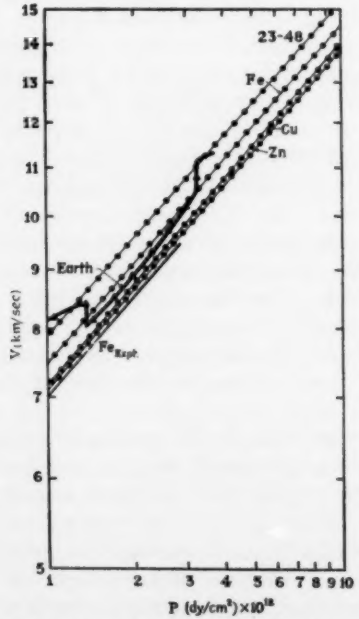


FIG. 4.—Velocity of sound in the fluid state at core pressures.

values of the velocity in iron. The velocity of seismic waves in the core of the Earth is about 0·4 km/s higher than the experimental values for metallic iron. Quantum-mechanical velocity equations of state have also been drawn for materials of atomic number 23 and for copper and zinc. The difference in velocity between a material of atomic number 23 and iron, and in turn between iron and copper, is again about 0·4 km/s. If there are no corrections for temperature, it is highly likely that the velocity in the core of the Earth is more appropriate to the material having atomic number 23 than to iron.

## 5. Density equation of state of the core of the earth

From the experimental velocity data it seems highly likely that the Earth's core has a density less than that of iron. A similar conclusion can be drawn from

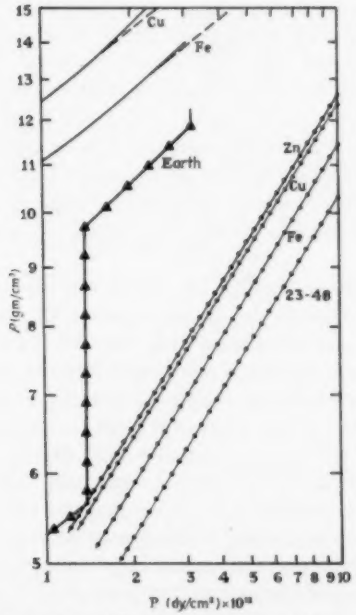


FIG. 5.—Equation of state for iron, copper, zinc and a hypothetical material of atomic number 23, atomic weight 48 in the core pressure range. The values derived from shock wave measurements (solid) are compared with those obtained from the Thomas–Fermi theory (closed circles). Bullen's density distribution is shown for comparison (triangles).

the density equation of state. The absolute zero experimental equations of state of the nine metals and the quantum-mechanical equations of state can be compared with the density equation of state for the Earth. This has been done in Figure 1. The portions of the curves in the pressure range appropriate to the core of the Earth have been extracted and the data replotted in Figure 5. The experimental data for iron and copper at the top of the graph show densities somewhat in excess of those for the Earth. If no corrections were applied, the density of iron would be of the order of about 1·8 g/cm³ larger than that for the Earth's core. The quantum-mechanical equations of state for zinc, copper, iron and a material of atomic number 23 (atomic weight 48) have been plotted and are found in the lower portion of

Figure 5. If no corrections were to be applied, again it would seem highly likely that the core of the Earth has an atomic number less than that of iron.

Strong (1959) has measured the change in the melting point of iron to a pressure of 96 000 atmospheres. Extrapolating his data and using Simon's equation, he obtains a minimum core temperature of 2340 °C. The derived value of Grüneisen's ratio is 0·4, in disagreement with the experimental result of Altshuler & others for core pressures. Strong's data are not inconsistent with a value of $\gamma = 1{\cdot}6$, if Simon's equation for the melting curve is used. An extrapolation of Strong's melting point curve using $\gamma = 1{\cdot}6$ leads to a melting point of iron at 1·4 Mb of 4 000 °C. If the core were pure iron, the temperature of the core would be in excess of 4 000 °C. The presence of alloying elements tends to reduce the melting point.

The density of solid iron at 0 °K derived from shock wave data must be corrected for thermal expansion and the volume change on melting before a meaningful comparison can be made with geophysical data. The thermal expansion of iron over the temperature range 20°–900 °C is $1{\cdot}5 \times 10^{-5}$/deg C. Birch (1952) estimates, on the basis of the classical theory of specific heat, the change in thermal expansion with compression. Applying his results to the thermal expansion of iron, we expect a reduction of about one half in the zero pressure value of $\alpha$ for an increase in pressure of a megabar. Adopting an average value of $1 \times 10^{-5}$/deg C for the thermal expansion of iron, we see that a temperature on the order of a few thousand degrees Celsius would result in a decrease in the room temperature density of iron of only a few per cent.

The change of volume on melting of iron at a pressure of one bar is $3 \pm 0{\cdot}5$ per cent (Gilvarry 1957) though the quoted uncertainty may be much too small. The volume change at core conditions also may be estimated from the Simon equation provided some assumption is made regarding the entropy of fusion. Since the latter quantity is most uncertain, it appears that the best guess is that a change in volume on melting of iron of 2 to 3 per cent might be expected at core conditions.

The combined effects of thermal expansion and melting on the volume of iron are such as to reduce the density by about 5 to 6 per cent. The discrepancy between solid iron at 0 °K and the density of the core is about 20 per cent. Though the corrections for thermal expansion and melting may be in error by a factor of two, it appears unlikely that these corrections are large enough to reduce the apparent discrepancy in the densities of the core and of pure iron. Furthermore, any nickel alloyed with iron would increase the discrepancy since the density of pure nickel is 8·6 g/cm³ at room conditions, and the density of nickel is greater than that of iron at extremely high pressures.

*Publication No.* 162
*Institute of Geophysics*
*University of California,*
*U.S.A.*
1959 *September.*

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# Pendulum Observations at Teddington, Singapore, Darwin and Melbourne in 1959

**J. E. Jackson**

(Received 1959 December 9)

*Summary*

Observations were taken with the Cambridge pendulum apparatus at Teddington, Singapore, Darwin and Melbourne in 1959 August and September and the results are reported in this paper.

In 1959 August and September the Cambridge pendulum apparatus was used in a series of measurements to determine the differences of gravity acceleration between the places mentioned in the title. The work was done at the suggestion of Dr George P. Woollard, who arranged for funds to be provided through the Woods Hole Oceanographic Institution. The inclusion of a visit to Tokyo in the itinerary was considered, but this was found to be impossible on account of lack of time.

After making the initial observations at Teddington on August 12 and 13, I left London Airport on August 15 and returned to Teddington on September 27. All travelling was done by commercial airlines. As before (Jackson 1959), the pendulums, manometers and quartz crystals were carried in hand, and the rest of the apparatus travelled as excess baggage except on the last flight, Singapore to London, when it was sent as air freight to minimize expense.

I was accompanied throughout the trip by Mr J. A. Paterson, a student of geodesy, who gave valuable assistance in technical and other ways: I have pleasure in acknowledging his contribution to the successful outcome of the work.

Each station was occupied twice, including Melbourne, where two complete sets of observations were taken, to make the outward and return series independent of each other. The apparatus and observational procedure were practically the same as for last year's observations in South America (Jackson 1959). Three new manometers in wide tubes were taken, and a small attaché case was fitted up for carrying these items together with the quartz crystals and the test magnetometer. A dip-circle also was included in the gear. Fortunately the heavy hand-operated vacuum pump could be left behind, because arrangements were made to have power-operated pumps available at all the stations. In the event, owing to some mischances, three pumps appeared at Darwin!

At the three overseas stations, the one-second interval pulses from transmitter WWVH were used for finding the true rate of the crystal clock. The rate remained remarkably steady over the whole journey, within the range 26 to $34 \times 10^{-8}$: this was fortunate, because the radio receiver was out of action during the first visit to Darwin and a rate of $32 \times 10^{-8}$ was assumed in the computation of pendulum

periods. Since the rate at the previous station was $34 \times 10^{-8}$ and at the next station was $30 \times 10^{-8}$, it seems unlikely that the assumed value can have had any significant error.

As before, a local pulse signal on a cable was available at Teddington.

The manometer was tested against a standard barometer at Teddington at the beginning and at the end of the period of observations: it was found to have acquired an error equivalent to about 0·8 mgal during use. The uncorrected pendulum periods give no indication that this error was acquired suddenly at any time, and the manometer readings were therefore adjusted on the assumption of a steady acquisition of the error.

Tests for magnetization of the pendulums were made in the usual way at all stations. Very little degaussing was necessary, except on the return to Teddington, where all pendulums were found to have acquired large magnetic moments.

We are greatly indebted, as usual, to many persons without whose co-operation these observations could not have been carried out. The arrangements made for provision of accumulators, pumps, and dark room facilities, and for numerous journeys between airports, hotels and gravity sites, were very much appreciated. At Singapore, Mr M. N. Menon, Chief Surveyor, and several members of his Staff gave much of their time to ensuring the progress of the work. We are indebted also to Dr R. Wikkramatileke, Head of the Department of Geography, and his Staff, who were deprived of the use of their map room for several days, and to Professor K. M. Gatha for permission to use the dark room in the Department of Physics.

The initial arrangements regarding the stations in Australia were made by Dr J. M. Rayner, Director of the Bureau of Mineral Resources, and Mr R. F. Thyer, Chief Geophysicist. At Darwin, Mr C. E. Prichard and Mr G. F. Clarke and other officers of the Bureau of Mineral Resources gave every possible assistance, and we are particularly grateful for the attention to the radio receiver which went out of action there. At Melbourne, we were looked after by Mr J. Goodspeed and others, and here, too, we received valuable technical help in the restoration of the radio receiver and the repair of one of the instrument boxes. At Teddington we had the cooperation of Dr A. H. Cook who arranged the usual facilities for transport and Customs examination of the apparatus: we are grateful also to the member of the National Physical Laboratory staff who cleaned and readjusted the test magnetometer before observations started, and to the member who gave necessary expert attention to the crystal clock at the return visit.

The corrected periods of the pendulums are listed in Table 1. At each occupation, the customary four pairs of pendulums were each swung twice. The thirty-two differences in the pairs of determinations of mean period have a r.m.s. value of $10 \times 10^{-8}$s and maximum $21 \times 10^{-8}$s. This indicates that a gravity difference calculated from one pair of pendulums should have a standard deviation of 0·57 mgal—the equivalent of $14 \times 10^{-8}$s in period. Thus the mean of eight such gravity differences would have a standard deviation of about 0·1 mgal. This is practically the same as the corresponding estimate for the results reported in the previous paper.

Differences between results of first and second occupations at the stations are listed in Table 2.

Computed gravity differences are given in Table 3, and they appear to be satisfactory. Further consideration of their accuracy is deferred, however, for the reasons given in my previous report.

## Table 1

*Corrected periods of pendulums in seconds*

| Pendulum: | Outward journey I.A | I.B | I.C | | Return journey I.A | I.B | I.C |
|---|---|---|---|---|---|---|---|
| 0.506 | 24871 | 25067 | | | 24839 | 25053 | |
| | 24851 | 25074 | | Teddington | 24815 | 25073 | |
| | 24862 | | 24763 | Aug. 12 Oct. 1, 2 | 24849 | | 24748 |
| | 24830 | | 24785 | | 24811 | | 24761 |
| 0·507 | 05390 | 05624 | | | 05361 | 05610 | |
| | 05380 | 05606 | | Singapore | 05343 | 05599 | |
| | 05396 | | 05298 | Aug. 19 Sept. 19 | 05368 | | 05300 |
| | 05369 | | 05287 | | 05342 | | 05304 |
| 0·506 | 99351 | 99583 | | | 99374 | 99583 | |
| | 99358 | 99561 | | Darwin | 99370 | 99589 | |
| | 99338 | | 99270 | Aug. 28 Sept. 14 | 99356 | | 99280 |
| | 99349 | | 99270 | | 99314 | | 99284 |
| 0·506 | 56288 | 56498 | | | 56276 | 56494 | |
| | 56268 | 56491 | | Melbourne | 56271 | 56494 | |
| | 56259 | | 56189 | Sept. 2 Sept. 8 | 56274 | | 56207 |
| | 56267 | | 56175 | | 56262 | | 56191 |
| Pendulum: | VI.A | VI.B | VI.C | | VI.A | VI.B | VI.C |
| 0·505 | 46289 | 46396 | | | 46309 | 46383 | |
| | 46282 | 46403 | | Teddington | 46287 | 46399 | |
| | | 46398 | 46208 | Aug. 13 Oct. 2, 3 | | 46434 | 46219 |
| | | 46434 | 46214 | | | 46446 | 46216 |
| 0·506 | 26719 | 26785 | | | 26738 | 26794 | |
| | 26711 | 26796 | | Singapore | 26722 | 26798 | |
| | | 26829 | 26626 | Aug. 20 Sept. 20 | | 26837 | 26619 |
| | | 26829 | 26616 | | | 26833 | 26612 |
| 0·506 | 20697 | 20796 | | | 20719 | 20744 | |
| | 20703 | 20788 | | Darwin | 20687 | 20774 | |
| | | 20809 | 20596 | Aug. 29 Sept. 15 | | 20824 | 20595 |
| | | 20799 | 20577 | | | 20794 | 20601 |
| 0·505 | 77693 | 77776 | | | 77674 | 77759 | |
| | 77669 | 77763 | | Melbourne | 77671 | 77762 | |
| | | 77805 | 77579 | Sept. 3 Sept. 9 | | 77798 | 77589 |
| | | 77802 | 77579 | | | 77803 | 77578 |

**Table 2**

*Differences between mean periods of pendulum pairs at first and second occupations: in $10^{-8}$ seconds*

(25 is equivalent to 1 mgal)

| | Pendulum pair: I.AB | I.AC | VI.AB | VI.BC |
|---|---|---|---|---|
| Melbourne | − 2 | +12 | − 8 | + 1 |
| Darwin | +16 | + 1 | −15 | + 8 |
| Singapore | −22 | − 8 | +10 | 0 |
| Teddington | −21 | −18 | + 2 | +16 |

**Table 3**

*Computed differences of gravity acceleration in mgals*

Upper figure—outward journey. Lower figure—return journey.

| | Pendulum pair I.AB | I.AC | VI.AB | VI.BC |
|---|---|---|---|---|
| Teddington | | | | |
| | −3114·3 | 3114·0 | 3114·4 | 3114·4 |
| | 3114·3 | 3114·4 | 3114·7 | 3113·8 |
| Singapore | | | | |
| | + 232·9 | 232·7 | 232·1 | 233·0 |
| | 231·5 | 232·3 | 233·1 | 232·7 |
| Darwin | | | | |
| | +1664·6 | 1664·9 | 1665·0 | 1664·3 |
| | 1665·2 | 1664·4 | 1664·7 | 1664·6 |
| Melbourne | | | | |

The stations occupied were:

*Teddington:* The British National Gravity Reference Station, National Physical Laboratory, Standards Division, Room 11.

*Singapore:* Map Room on the ground floor of the Department of Geography, University of Malaya. At the first occupation, site marked $X_1$ in room F, was used as this is the site occupied by the Japanese party in 1957. At the second occupation, site $X_2$ in room H, was used so as to cause less inconvenience to the staff of the Geography Department, and the difference of gravity between these two positions must be quite inappreciable. (The small air-conditioners fitted to windows in these rooms were left running.)

*Darwin:* Radioactivity assay room in the office of the Bureau of Mineral Resources. (The air conditioning machine in this room was switched off during observations: the machine in the adjoining room was left running.)

*Melbourne:* A small room on the ground floor just inside the central entrance of the Geophysical Laboratory at Footscray. (The local magnetic field here is disturbed by the passage of electric trams on the road outside: the usual tests of pendulum magnetization could not be carried out till after the trams stopped running at 1 a.m.) The same site was occupied by the Japanese party.

Diagrams of the overseas sites are given in Figure 1.

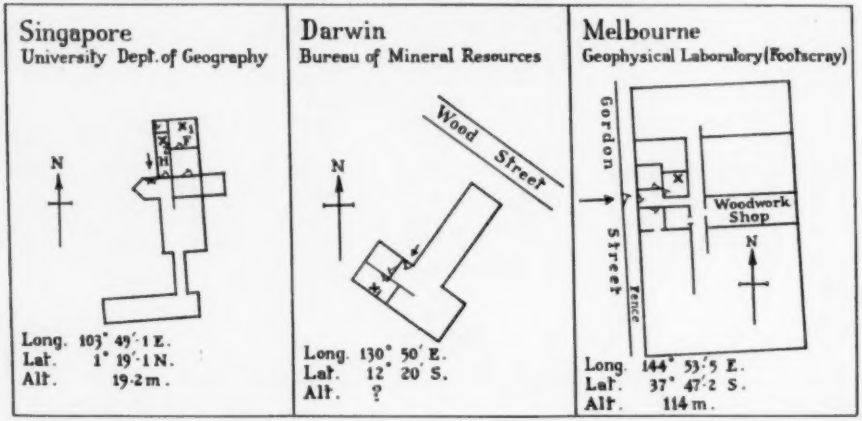


FIG. 1.—Location diagrams of foreign stations. Pendulum site: ×.

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1959 *December* 2

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# Notes on Progress in Geophysics

## The Figure of the Earth
## Its departure from an exact spheroid

**Brig. G. Bomford**

*Summary*

It has long been known that the figure of the Earth, defined as that of the Geoid or sea-level equipotential surface, is not exactly an oblate spheroid, and it is recognized that the aim of the geodesist is not only to determine the dimensions of a "best-fitting" spheroid, but to map the true form of the geoid by contoured charts showing its departure from some defined spheroidal datum surface. A. H. Cook (1959) describes recent progress in the determination of the low-degree harmonic constants in the Earth's gravitational field, and thence in the deduction of the dimensions of the best-fitting spheroid. The present paper describes recent progress in the measurement of the irregular departures of the geoid from any spheroid, with special reference to their direct measurement by the method of geoidal or "astro-geodetic" section.

### 1. Introduction

When the figure of the Earth is described as being approximately an oblate spheroid, the surface referred to is not the ground-level surface, but the geoid: defined as that equipotential surface of the Earth's gravitation and rotation, which most nearly coincides with mean water level in the open ocean. Slight ambiguities due to mean water level not necessarily being an equipotential surface at all, and to periodic variations in its shape, are unlikely to exceed 1 metre, and in this context they can be ignored.

For more than a century it has been realized that the form of the geoid differs measurably from that of any oblate spheroid. Such departures can be recorded in three different ways, each roughly corresponding to one of the three ways in which the irregularities of the surface can be measured. These are:

(a) The *form* of the geoid may be described geometrically as so many metres ($N$) above or below some arbitrarily chosen geometrical reference surface, which will most naturally be an oblate spheroid selected so as to approximate closely to the geoidal form. This method of record corresponds to the direct measurement of $N$ by geoidal section. (Section 5.)

(b) The observed value of *gravity* $g$ may be compared with the value $\gamma$ which would be expected if the Earth was indeed a spheroid of known flattening, with or without allowance for the visible topographical irregularities and

hypothetical isostatic compensation. The variations of $(g-\gamma)$ or $\Delta g$ are related to $N$, the anomaly of form, by Stokes's well-known theorem. (Section 4.)

(c) The anomalous variations in the *potential* of the Earth's attraction and rotation ($\Delta V = N \times g$, where $g$ is any approximate value of $g$) may be recorded. These may be got to some extent from observations of the orbits of artificial near-earth satellites (Section 3), or differentially as between one point and another near it, by direct calculation if the form of the ground and the densities of the rock are known.

It is repeated that these three anomalies—of form, gravity and potential, are three manifestations of the same phenomenon. Given one in sufficient detail all over the Earth, we know the other two, apart from an initial constant, and vice versa.

Each of these three forms of the anomaly may be numerically recorded in two different ways:

(i) Analytically by a series of spherical harmonics for $r$ or $N$, $g$ or $\Delta g$, $V$ or $\Delta V$ according to whether the form, attraction or potential of the standard Earth is or is not excluded.

or (ii) Graphically by contoured charts of $N$, $\Delta g$ or $\Delta V$ shown with reference to a standard figure or formula.

Analytical representation by spherical harmonics has obvious theoretical advantages, but numerical analysis on these lines is impracticable beyond some quite low degree of harmonic, such as $n = 10$* which involves 121 unknown constants, or conceivably $n = 100$ with 10 201 constants. Even then, seeing that the 100th degree sectorial harmonic has an equatorial wave-length of 260 miles, only the Earth's major geodetic features would be shown by such a labour, while such striking features as the great troughs of negative anomaly, which are only 100 km wide, would be smoothed out and lost. It follows that when $N$, $\Delta g$ or $\Delta V$ are known in any detail the graphical method is the only one capable of representing them adequately, and theorems involving spherical harmonics must either be put into such a form that results can be obtained by quadrature, as Stokes did for his expression for $N$ (given $\Delta g$), or recourse must be had to some approximate generalizations such as "If the mass of a series of parallel mountains can locally be represented by a skin-density of (spherical harmonic) degree $n$, the resulting $\Delta g/g$ will be $(n-1)$ times the corresponding $N/R$".

## 2. The reference figure or formula

The reference figure above which $N$ is recorded to give the form of the geoid, or the standard formula with which $g$ or $V$ are to be compared to give $\Delta g$ or $\Delta V$, may take one of two forms:

(a) For $N$, an exact oblate-spheroidal surface whose equation expressed in Legendre's functions contains $P_2$, a small $P_4$ and a quite negligible $P_6$, (Cook 1959, p. 226). And similarly $\Delta g$ and $\Delta V$ may be reckoned from a

* Prey (1922) has analysed the form of the Earth's solid surface to the 16th degree spherical harmonic, and Vening Meinesz (1959) to the 31st. Zhongolovitch (1952) has analysed the gravity field to the 8th degree, and has deduced the corresponding geoidal form. Kaula (1959) has done the same with later data.

formula giving the values of $g$ and $V$ on the surface of a body of perfect oblate-spheroidal form, filled with matter so distributed in density that its surface is an equipotential surface of its own attraction and rotation.

or (b) Since an oblate spheroid filled with *liquid* whose density increases towards the centre cannot be an equipotential surface as above, a figure with a slightly different $P_4$ term (and no doubt a completely different $P_6$) has sometimes been proposed for $\Delta g$, such that allowing for the heavy centre its surface could be an exact equipotential. In gravity the difference in the $P_4$ term nowhere amounts to more than 2 mgal, which (for $N$) corresponds to 3 m in the bounding surface (Bullard 1948). Such a standard has seldom, if ever, been adopted for $N$ or $V$. From the geophysical point of view there is advantage in adopting a strengthless body as a standard object from which to reckon anomalies, but as the difference between it and the perfect spheroid is so small, it is more convenient to adopt the latter, as does the International gravity formula of 1930.

It is to be remarked that the dimensions and position of the centre of a reference spheroid and the constants of a standard gravity formula are arbitrary. The surface of the geoid can be correctly represented by values of $N$ which vary according to the spheroid adopted, but there is obvious convenience and better clarity of representation if the spheroid is nearly the one of "best fit" as regards its dimensions and if its centre is as close as possible to the Earth's centre of gravity. And the determination of the best-fitting constants by each of the three methods is a worthy objective, both in itself and from the point of view of comparing and combining their several results.

### 3. Direct determination of the potential

The geoidal potential anomaly (and hence the geoidal rise $\Delta V/g$) caused by an abnormal body of any specified size, shape and density can readily be calculated by elementary potential theory, aided by tables and quadrature, if necessary. As an example of orders of magnitude, a mountain which can be generalized as a vertical cylinder $\frac{1}{2}$ mile high and 10 miles in diameter will raise the geoid by about 2 feet (and it will change $g$ at ground level in its centre by 90 mgal) if it is not compensated by local isostasy, while compensation on Hayford's system makes little difference. For a plateau of diameter 200 miles and height $\frac{1}{2}$ mile, these figures are about 50 feet and 90 mgal (uncompensated), which would be reduced to 13 feet and 30 mgal if Hayford compensation existed below. As another example, under a series of parallel mountain ranges averaging 2000 feet high above the intervening valleys and 100 miles apart from crest to crest (uncompensated), the geoidal undulations would be $\pm 2{\cdot}9$ feet. (Bomford 1952, p. 322). Such examples help to show the magnitude of the mass anomalies required to produce such irregularities as are actually found, and to indicate whether they can be caused by visible topography or demand the presence of invisible abnormalities of density in addition.

The direct determination of potential anomalies from the motion of a close satellite has been dealt with by Cook (1959), to which little new information can yet be added. De Graaff-Hunter, King-Hele and others have found expressions for the changes in the orbit when the expression for the gravitational field involves a series of Legendre terms of any degree, and the rate of regression of the node of

every satellite with a different inclination between the equator and the plane of its orbit produces an equation among the even coefficients $J_2, J_4, J_6$ etc., on the assumption that those not solved for are zero. The odd-numbered $J$'s can also be determined with rather less certainty. Cook (1959) gives the best results, for $J_2$ to $J_6$, so far available. In course of time reliable values may be obtainable as far as (say) $J_{10}$, if the subsequent ones can justifiably be neglected. One feels little doubt that $J_2$ will eventually be determined with an accuracy corresponding to 0·1 in the inverse of the flattening, but one may feel uncertain whether or not the series of $J$'s converges rapidly enough for the values obtained for $J_5$ and later to be significant, when all after (say) $J_{10}$ are neglected. (See Kaula (1959) p. 51–52, where the "normalized"* gravity coefficients for $n = 5$ average 0·62 mgal, while for $n = 8$ the figure is 0·52). Those for geoidal height, with an additional factor of $1/(n-1)$, of course converge more quickly, namely from 1·01 m to 0·48.

## 4. The determination of the geoid from gravity observations

### 4.1 *Stokes's theorem and* $g-\gamma_A$

During the last 10 years the number of observed gravity stations has greatly increased and their distribution has improved, and with the introduction of gravity meters suitable for use on surface ships rapid further progress is to be hoped for.

Theory has also progressed, in that the long-standing trouble that the geoid is not a bounding surface, and that gravity observations on land are not made at geoid level but at a considerable distance above it, has been clarified by de Graaff-Hunter (1958), who shows that the anomaly of potential at ground level at any point is $Ng$, where $N$ is correctly given by Stokes's integral if $\Delta g$ is taken as $(g-\gamma_A)$, $g$ being gravity at ground level, and $\gamma_A$ is standard gravity $\gamma_0 - 2gh/R$, in which $R$ is the mean spheroidal (not geoidal) radius of curvature. Strictly, $g$ should receive an orographical correction in mountainous country, but the total effect on $N$ will be small, especially when $N$ is computed at a site remote from mountains. This point is met satisfactorily if de Graaff-Hunter's model Earth is employed. (See Section 4.2.)

From this point of view charts illustrating variations of gravity should then show $g-\gamma_A$, whose mean value must from them be assessed over each area used in the quadrature of Stokes's integral. The difficulty is that in hilly country $g-\gamma_A$ is extremely variable, and over small areas 1 mgal contours of $g-\gamma_A$ will correspond closely to 30-foot contours of height. Average values may thus be difficult to estimate. The trouble may be remedied by supposing the topography to be smoothed out laterally or otherwise moved or removed. Any such supposition changes $g$ at every observed point by an amount which can be more or less laboriously computed, but which results in a smoothly-varying anomaly. The movement of mass also changes the potential at the point where $N$ is required, and Stokes's formula consequently produces a "co-geoid" instead of the geoid itself. Here also, the separation between geoid and co-geoid is computable (with more or less labour) and the form of the geoid is determinate.

### 4.2 *Smoothing observed values of the anomaly*

The most obvious method is to suppose the topography above sea-level and the oceanic defects below it removed to an infinite distance. As a method of

* These are "normalized" coefficients, and equal the root-mean-square value of the term over the whole sphere. Maximum values may be 2 or 3 times as great.

smoothing this is effective, but the changes in $g$ and $N$ are large, and unnecessarily laborious to compute. The change in $N$ may be several hundred metres. The system is not a practicable one for use with Stokes's integral, although practical geophysicists find the old form of "Bouguer anomaly", $g-\gamma_B$,* very useful for charts used for geophysical prospecting.

As an alternative, the topography may be assumed to be moved to corresponding points inside the geoid (thereby making the latter a bounding surface), and as before calculating the effects on $g$ and $N$. If the system adopted is well chosen, these effects are now smaller, and provided the depth to which the masses are moved is reasonably large, such as 20 miles or more, the smoothing is equally effective. If the outside masses and oceanic defects are moved to the points supposed to be occupied by their compensation on Hayford's or any other isostatic hypothesis, as they legitimately may be without in any way assuming the truth of the hypothesis, isostatic reductions such as may have been computed for other purposes may be taken as they stand to give the change in $g$. The change in $N$ may (on Hayford's system) amount to 30 m, which is not negligible and must be calculated.

De Graaff-Hunter (1958) has proposed a model Earth formed from the actual Earth by lateral movements of masses to give a smoothed topography (height $H$), with slopes not exceeding one in 100. At the surface of the model Earth the gravity anomaly appropriate for use with Stokes's theorem is

$$g-\gamma_E = g-\gamma_A+A(H-h)+O$$

in which $A$ is the attraction of an infinite plate of normal surface density, 1 mgal per 29·4 feet; $h$ is the height of the actual station; and $O$ is the Orographical or "Terrain" correction, namely the vertical attraction at the station* of the Earth's actual topography, less that of its condensation on the equipotential surface through the station.† In practice it is not necessary to carry the computation to beyond 100 km from the station, and Earth's curvature can be ignored.

The change in $N$, as between the model and actual Earths, nowhere exceeds 7 m, and for most purposes may be ignored. The anomaly $g-\gamma_E$ runs as smoothly as Hayford's or other isostatic anomaly $g-\gamma_C$.

From the point of view of computing the form of the geoid, the position then is that except in flat plains $g-\gamma_A$ is too variable: $g-\gamma_B$ is useless: $g-\gamma_C$ (isostatic) is permissible provided the resulting co-geoid is corrected to give the geoid; and $g-\gamma_E$, once charts of the smoothed Earth have been produced, will give the geoidal form direct with minimum labour in computing the changes in $g$ and $N$.

In many parts of the world sufficient gravity data exist for excellent $g$-anomaly charts to be drawn as above, on the scale of 1/1 000 000 or larger, but their preparation has been held up by the great labour of the isostatic reduction, while the de Graaff-Hunter model has only recently been formulated. Large-scale published charts are mostly of $g-\gamma_B$, as favoured by geologists and mineral prospectors. Examples are:

* In what may be called "the old Bouguer anomaly" the topography around the station is assumed to be made up of an infinite plane plateau or plate as thick as the station height, whose attraction is about $3\gamma h/4R$, on which are imposed the local topographical irregularities. Earth's curvature being ignored, the attraction of the latter (known as the *Orographical correction*) is easy to compute. This system defeats clear thinking, but has practical use. The correctly computed topographical effect, allowing for curvature and comprising the whole Earth, is troublesome to compute.

† At actual Earth level $h$.

(a) Great Britain. A 1/253 440 series of $(g-\gamma_B)$ is in preparation. Only 3 sheets out of 33 have yet been issued, but charts of 60 or 70 per cent of England exist in manuscript or other published form.
(b) France. Carte Gravimétrique, 1951. $(g-\gamma_B)$. 1/1 000 000.
(c) France. Carte Gravimétrique, 1954. $(g-\gamma_C)$ (Airy). 1/1 000 000.
(d) Germany. (i) Karte der Bouguer-Isoanomalen. 1/1 000 000 von Westdeutschland, 1957.
(ii) Karte der mittleren Freiluftanomalien von Westdeutschland, 1959. Deut. Geod. Kommission Nr. 46, Teil II.

4.3 *Generalization from few stations*

For use with Stokes's integral, average values of $(g-\gamma_A)$ or $(g-\gamma_C)$ or $(g-\gamma_E)$ are required over finite areas such as 1° or 10° squares, with a detailed contoured chart of a small area surrounding the place where $N$ is being computed. In most 1° or 10° squares data are insufficient for a contoured gravity map, and at first sight, the degree-square or 10-degree-square mean can only be got by averaging the available values of the anomaly. But since $(g-\gamma_A)$, and to a lesser extent $(g-\gamma_C)$ and $(g-\gamma_E)$, over any small and hilly area is strongly correlated with height, it is desirable to allow for this correlation by fitting a straight line to a plot of $(g-\gamma)$ against height of station and recording the value of $(g-\gamma)$ corresponding to the average height of the square, as was done for $(g-\gamma_A)$ by Jeffreys (1941 and 1948). This anomaly even shows some correlation with height over $10^\circ \times 10^\circ$ squares, but less strongly than over the small areas. It is open to question whether this is worth doing with $(g-\gamma_C)$ or $g-\gamma_E$.

Zhongolovitch (1952) has devised a system of equal-area "10-degree" squares, which have also been used by Kaula (1959). There is much to be said for their adoption internationally for the recording of mean anomalies and average heights. The system can easily be extended to give equal-area "1-degree squares".

## 5. Astro-geodetic section

The third method of geoid determination, the oldest and most direct, is the astro-geodetic. It is not necessarily better than the other methods, but it produces results which cannot be ignored. Recent literature contains few references to current methods, and the larger part of the present paper is therefore given to a description of it.

5.1 *History*

The earliest measurement of a pair of meridian arcs may have been thought to have determined the "axes of the spheroid", but over 100 years ago solutions were being made from a number of arcs to find a "best-fitting spheroid", and it was realized that incompatibilities arose from "local attraction"* (deviation of the vertical), or distortion of the sea-level surface, as much as from errors of measurement. Hayford (1909) integrated these deviations in the United States to determine the form of the geoid there, and in 1922 de Graaff-Hunter (1928 Chart I) did the same in India. Both showed the form of the geoid by contoured charts giving its height above the reference spheroid used for their national triangulation

* "Local attraction", suggesting that the anomalous direction of gravity is caused only by local anomalies of mass, was a bad name for what is now called "Deviation of the Vertical".

systems. Later, de Graaff-Hunter introduced a new method of finding the best-fitting spheroid, namely by finding the dimensions of the spheroid, and coordinates for its centre, which would best fit the geoid or co-geoid* in the U.S. and in India. (J. de Graaff-Hunter 1929, p. 181 and 193).

These early geoid charts were based on such values of the deviation as had been provided by the industry and foresight of their compilers' predecessors: numerous values derived from astronomical latitudes, a few from longitudes, and many, but not very reliable, values derived from azimuths. Observed deviations were apt to be 100 miles apart, and the approximation $N_{AB} = \frac{1}{2}l(\zeta''_A + \zeta''_B) \sin 1''$ for the difference of $N$ between stations A and B situated $l$ miles apart, was often quite unjustifiable. In India there followed a programme of observing deviations at intervals of 10–15 miles along cross-country section lines, and by 1940 9000 miles of such section had been observed, a distance which has since been increased to 11 500 miles (in what is now India, Pakistan and Burma). These sections cover the whole country from Iran to Indo-China and southern Thailand. Details have been given in the annual Geodetic Reports of the Survey of India.

Before 1940 similar sections had been observed in Switzerland, across the Alps, 190 km at 3 or 4 km intervals (Niethammer 1939): in Argentina in 1926, 200 miles in Entre Rios: and in Germany some sections in Hesse, Wurtenburg and the Harz mountains as long ago as in 1895–1910. The last two items crossed no very notable geographical or geophysical features, showed no startling results, and aroused less general interest than they deserved.

Since 1948 the observation of geoidal sections has received more world-wide attention. A main section line from southern Burma through India and Turkey to Scotland is probably complete, except for some computations in Iran and some gap-filling observations in Turkey and Greece (recent reports are lacking). Another main section line extends from northern Finland to the Mediterranean, whence it continues (with two or three rather wide gaps) down the 30th meridian to South Africa. Russian work has provided a geoidal section from Europe to Vladivostock, but no internal detail has been published. (Dubovsky 1939.) All the above have been expressed in terms of a common spheroid and datum, Potsdam 1950 International Spheroid. The geoid in the United States (Fischer 1957) is being reinforced by an east–west section in latitude 35° N, which is nearly complete. Unpublished results of radar tri-lateration between Norway and Canada enable the United States geoid to be expressed in European terms, except that the value of $N$ in European terms at any single North American point is still lacking. (See Section 5.4(c) and (d) below).

### 5.2 *Method*

The basic principles by which astro-geodetic section determines the form of the geoid are not always clearly understood. As an illustration the method for an ideal straight meridian section, not involving deviations in longitude will be given. Needless to say, sections need not be meridional or straight.

The two axes of the reference spheroid are arbitrarily defined, and the minor axis is defined to be parallel to the Earth's mean axis of rotation† which points towards the celestial pole, which is recognizable from the Earth's surface. An

* The co-geoid is derived by integrating Hayford anomalies of deviation.

† Polar wandering is not mentioned here, but observed astro lat. and long. should be corrected if possible. The axis of the reference spheroid must so far as possible remain fixed in relation to the body of the Earth's solid surface.

arbitrary one-to-one correspondence is set up between points $p$ on the ground, and points $P$ on the spheroid, namely that if $p$ lies on the spheroidal normal of $P$, the inclination of the normal to the equator is the geodetic latitude of both. At a certain point $p_0$ known as the origin, the height of $p_0$ above $P_0'$ (on the geoid) is measured by spirit-levelling and the separation of geoid and spheroid, $N_0 = P_0'P_0$, is arbitrarily defined as zero or otherwise. And the *geodetic* latitude of $p_0$ and $P_0$ (essentially equal) is defined to be $\phi_0$, which may or may not be equal to $\phi_{Ast}$($C\,p_0\,T_0$ in Figure 1), the elevation of the celestial Pole above the tangent to

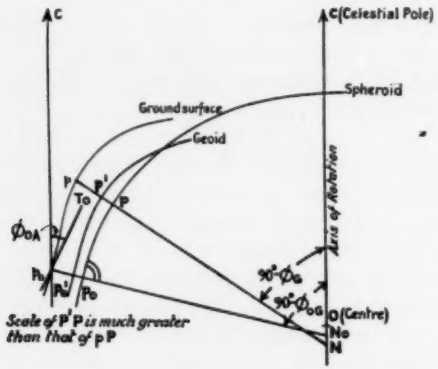


FIG. 1

the level surface at $p_0$, which is known as the *astronomical* latitude of $p_0$. The difference (astro minus geodetic latitude) or $\phi_A - \phi_G$ or $(A-G)$Ast is known as $\xi_0$, the deviation of the vertical (in meridian). A similar process* defines geodetic longitude at $p_0$ and completes the location of the centre of the spheroid with reference to $p_0$ and to the directions $p_0C$ and the normal to the equipotential surface passing through $p_0$.

Now let triangulation or traverse, and spirit levelling, survey processes, which amount to the direct measurement of linear distance on the surface of the spheroid and height above the geoid, proceed from $p_0$ to any other ground station $p$. If these are correctly observed and computed (Bomford 1952, p. 73–76) this will measure the spheroidal distance $P_0P$ and the geodetic latitude of $P$ and $p$. Astronomical latitude is also observed at $p$, hence $\xi$ the deviation at $p$, and so on at intervals of 10–20 miles. We now make two assumptions:

(a) that the geoidal tangent at every $P'$ is parallel to the tangent to the ground level equipotential surface at $p$, except for the small inclination which would be correct if all equipotential surfaces were undisturbed concentric spheroids; and

(b) that the change in $N$, $(PP' - P_0P_0')$, or $N_{AB}$ between two stations A and B separated by a distance $l$, is $\frac{1}{2}l(\xi_A+\xi_B)\sin 1''$, where $\xi_A$ and $\xi_B$ are deviations (in this case in meridian) along AB.

Starting from the arbitrary separation at $P_0$, the height of the geoid above the spheroid is then determinate, apart from errors of observation and of these two

* Details not given here. See Bomford (1952) p. 68–72.

assumptions (see Section 5.3). And the form of the geoid, considered as a surface in space is correctly determined whatever values may have been accepted for the arbitrary constants involved in the definition of the spheroid. Other national surveys, independently based on other spheroids or origins, will not of course have the same values of $\phi$, $\lambda$ and $N$ at common points, but the existence of a single common point establishes the three equations necessary for determining the arbitrary $\phi_0$, $\lambda_0$ and $N_0$ of one country in terms of those of the other, so that unification on one spheroid and datum becomes possible.

## 5.3 *Accuracy*

The accuracy of the two assumptions made in 5.2 must be examined.

As regards (a). Neglect of the theoretical correction, which is systematic in sign, involves an error of 5 m in $N$ in a section from equator to pole at an average height of 1 000 m above sea-level: not a serious matter, but one that can very easily be avoided. The error due to topography is random in sign.

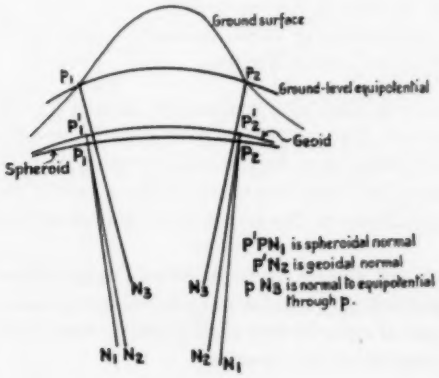


FIG. 2

It is clear that the inward deviation towards a mountain will be greater on its flanks at $p_1$ and $p_2$ in Figure 2 than at $P_1'$ and $P_2'$ at geoid level. The difference can be quite large, such as 5″, as has been calculated at the sub-Himalayan station of Jharipani near Dehra Dun, but the effect on a geoidal section will largely cancel as between one side of a hill and another. Also, deviation stations deliberately avoid hill flanks and are situated either on hill-tops or valley-centres, where the effect is unlikely to excede 1″, and to be random as between one station and another 15 miles away, except in such a situation as on the south side of the Himalayas, where an average (but far from steep) slope of constant sign may persist for four or five intervals. More positive evidence is as follows:

(i) In the Swiss geoidal section across the Alps with a station spacing of 3 to 4 km, the section has been computed with and without allowance for the "curvature of the vertical", which has been computed from the known topography at each station. (Niethammer 1939). The difference of geoidal rise is 0·4 m in the centre, falling to 0·2 m at the end of the 190 km section with a suggestion of further reduction if the section was continued to the Italian plain.

(ii) In so far as the error is random it is included in the total random error investigated with the second assumption below, and presumably accounts for no more than a small part of it, for the difference in deviation as between ground level and sea level is clearly likely to be, on average, materially smaller than the total deviation at either.

As regards (b), the error in the assumption that $N_{AB} = \frac{1}{2}l(\xi_A+\xi_B)\sin 1''$ can be investigated in several ways, such as by abstracting the closing errors round the three sides of triangles (Olander 1954), or by comparing the observed value of the deviation at one station, in a straight section, with the mean of those on either side of it, for which the very closely-spaced stations of the Swiss Alpine section provide valuable data for a mountainous region. Any figures derived for the error by either of these methods will also include random error in the astronomical observations, and in the assumption that sea-level and ground level equipotential surfaces are parallel. Various formulae for the total probable error of $N$ from these random causes are:

(i) $\pm 0{\cdot}008l\sqrt{(L)}$ metres in Alpine country,
(ii) $\pm 0{\cdot}00035l\sqrt{(L)}$ metres in ex-Himalayan India,
(iii) $\pm 0{\cdot}00024l\sqrt{(L)}$ metres in Finland,

where the total distance is $L$ km and stations are $l$ km apart. If $L = 4\,000$ km and $l = 25$ km, the three formulae give a total of $\pm 2{\cdot}5$, $0{\cdot}5$ and $0{\cdot}3$ m respectively, all reasonably negligible especially as Alpine country can generally be avoided.

More serious sources of error are those in the geodetic triangulation, and on account of personal equation in the longitude observation when sections are not meridional.

The above is discussed more fully in Bomford (1960) where the probable error of $N$ after 4 000 km is given as $\pm 2$ m for personal equation in an east–west section, and $\pm 1{\cdot}7$ m on account of error in first class geodetic triangulation. The probable error of $N$ from all sources is then given as

$$\pm 1{\cdot}4\sqrt{(K)} \text{ metres}$$

after $1\,000K$ km with a station spacing of 25 km, provided all work is up to the usual specification in other respects. The closing errors of nine Indian circuits of 1 000 to 3 000 km are in good general accord with this formula.

### 5.4 *Expedients*

Where oceans or other difficulties enforce a wide spacing of astronomical stations, or where ragged values among large mountains show that the usual spacing is inadequate, various expedients may be resorted to, such as:

(a) The integration of Hayford deflection anomalies to give locally a section of the co-geoid, from which the geoid section is determinate. In rough country the Hayford anomalies run more smoothly than plain $(A-G)$.
(b) Vertical angles, relying on accurate computation of the refraction, can carry useful values of the deviation through short distances at high levels, and so provide a more closely-spaced section between astronomical stations (say) 20 miles apart.
(c) Local gravity observations may aid interpolation and integration in (a). At present probably the most practical aid to crossing a wide sea-gap.

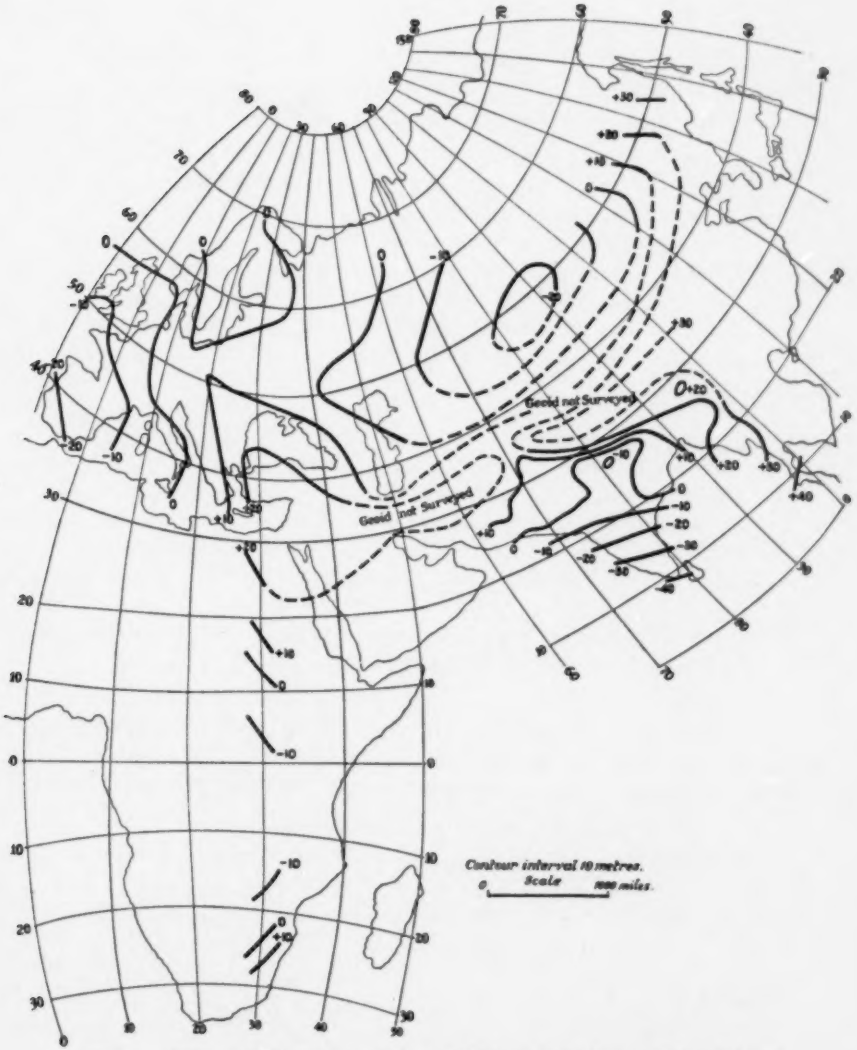


FIG. 3.—The geoid in Europe, Asia and Africa. Oxford 1959 Spheroid and datum:

$a = 6\,378\,201$ m. $f = 1/297{\cdot}65$.

(d) "Astro Triangulation", namely the photographing of rockets or satellites against a stellar background, may also be a possible means of carrying $N$ across a sea many hundreds of miles in width.

5.5 *Numerical results*

Values of the deviation of the vertical have been printed in innumerable publications but most of those which are sufficiently closely-spaced for integration to give the form of the geoid are contained in the following:

(i) International Association of Geodesy Lists 2 and 2(bis) presented at Toronto in 1957, give about 1800 European stations on the Potsdam datum. Compiled by J. J. Levallois.

(ii) Tome XX Fasc. 5 of the *Travaux* of the International Association of Geodesy (1960) will give 360 stations in Africa, also on the European Potsdam Datum. Bomford (1960).

(iii) Survey of India Technical Paper No. 9, 1955, compiled by B. L. Gulatee, gives 1070 deviations on the Indian Everest and "International" datums, with another 250 (less reliable) derived from azimuth observations. Additions will be found in annual Technical Reports.

(iv) Japanese deviations, on a local datum, are in No. 1 Report of the Japanese Geodetic Commission 1951, compiled by M. Torao, with Supplements, about 150 in all.

(v) For Canada, C. H. Ney (1952) gives 266 values which are supplemented by another 98 in the Canadian National report presented at Toronto in 1957. (Tome XX, 1960 of the *Travaux* of the A.I.G.). All on the North American Datum.

(vi) For the United States, J. A. Duerksen (1941), gives 737 stations and in a supplement in 1956 gives later stations also on the North American Datum.

Figure 3 shows the geoid in Europe, Asia and Africa, so far as it is at present known, as determined by geoidal section with reference to a single best-fitting spheroid and datum. Larger scale geoid charts have been published as below:

(a) Europe. G. Bomford. Bulletin Géodésique **42,** p. 45, 1956.

(b) India, Pakistan and Burma. Annual Geodetic Reports. Also Gulatee, item (iii) above.

(c) North America. I. Fischer. Presented to the A.I.G. at Toronto, 1957. And in Fischer (1959).

*Hainton Lodge,*
*Sutton Courtenay,*
*Berks.*

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# Palaeomagnetic Pole Positions, Part 1

## Pole numbers 1/1 to 1/144

**Compiled by E. Irving**

(Received 1959 October 12)

This list contains the pole determinations available up to 1959 September, with the exception of some results from Russia. Further parts will appear, approximately every half year, as new determinations are made. The Russian results will be given in a future issue. The results are presented as a table with 19 columns. The details given in originals has not always been sufficient to allow entries in all columns.

Polar identification numbers are entered in column 2. The numbers in each part are arranged approximately according to age. It is felt that this number, prefixed by the part number, will be useful for identification purposes and in figure illustrations, and also as a key to the original. For instance, the pole determination from the Rosehill Formation has the reference number **1/117**, the first number referring to Part 1 of the list and the second to the position within the part where a reference to the original may be found.

Information about the rock unit from which the pole determination has been made is given in columns 3 and 4. Its geological age is specified by these period symbols: Quaternary Q, Tertiary T, Cretaceous K, Jurassic J, Triassic Ʈ, Permian P, Carboniferous C, Devonian D, Silurian S, Ordovician O, Cambrian Ꞓ, and Pre-Cambrian pЄ. Subdivisions of periods from Cambrian to Cretaceous, where recognized, are denoted by u, m or l. Epochs within the Tertiary (Palaeocene, Eocene, etc.) are denoted by Tpa, Te, To, Tm and Tp. The Quaternary is divided into Recent Qr, and Pleistocene Qp. Subdivision of the Pre-Cambrian is not attempted. If it is thought that a rock formation was magnetized long after formation the probable "magnetic age" is entered in brackets after the geological age. Where possible, a distinction is made between a collecting *site* which is thought of as the particular place (say a quarry) from which a specimen is collected, and a collecting *locality* which refers to a more general area (for example, a canyon section) which may comprise several sites at different levels in a rock sequence. However, the variation in sampling schemes used by different workers does not always allow this distinction to be made.

The thickness of the rock unit which has been sampled is given in column 5 and in column 6 the spread of sampling localities (or sites) is indicated either as an area, or as a distance between extreme localities in those cases where the localities are few or are distributed along a zone rather than over an area. Pieces of rock, separately orientated, are referred to as *samples*, and from them one or more *specimens* are cut. The numbers of these samples and specimens are entered in columns 7 and 8. The distinction between samples and specimens has not always been made in originals, and in cases of doubt the entry is made in column 7.

Stability tests, where made, are noted in column 9. The tests are of five main types which are identified by these abbreviations; conglomerate *cg*, fold *fd* and igneous contact work *ct*, and thermal *th* and alternating field *ac* demagnetization. Remeasurements of specimens after storage, and deviation tests (deviations from the present Earth's field) are less critical and are not mentioned. Presumptive evidence relevant to stability is obtained from studies of the variation of saturation magnetization with temperature and of certain ratios such as the Koenigsberger ratio, but although important in other respects this evidence is indirect and is not mentioned here.

In most cases, denoted by *nt* in column 10, pole positions have been obtained from measurements of the direction of the natural remanent magnetization. Sometimes the specimens have been partially demagnetized to remove the secondary components and these cases are denoted by *ac* (alternating field demagnetization) or *th* (thermal demagnetization) in column 10.

Columns 11 and 12 give the proportions of positive and negative polarizations recorded. The convention is to regard the field as negative (positive) when the pole towards which the north (south) seeking magnetization points is in the present day northern hemisphere. These "polarity ratios" are approximate only.

In columns 13 to 15 estimates of the mean direction of magnetization are given (Fisher 1953, Watson & Irving 1957). The declination $D$ is reckoned clockwise east from geographic north, and $I$, the inclination, is regarded as negative if the direction is upwards and positive when below the horizontal. For rock units in which the beds are undisturbed the horizontal is the present day horizontal, but if the beds have been tilted the horizontal is assumed to be the bedding plane which, in the case of igneous rocks, is obtained from adjacent sediments. The error angle $\alpha$ is the half-angle of the cone whose axis is the direction $(D, I)$ and within which the true mean lies with a probability of 95 per cent.

In columns 16 to 18 the northern hemisphere co-ordinates of the poles are given. The errors $d\psi$ and $d\chi$ are the semi-axes of the elliptical error area round the pole at a probability of 95 per cent, $d\psi$ in the colatitude direction and $d\chi$ perpendicular to it. Polar errors are given only in those cases where it may be supposed that the secular variation has been meaned out. They may be readily obtained from the values given in columns 14 and 15. The poles have usually been obtained by computation but also sometimes by graphical methods. Reference to the original will be necessary to obtain such details. In some cases the pole positions were not calculated by the workers who made the observations, and this has been done either by others, to whom reference is made, or specially for this table. The latter cases are indicated by an asterisk in column 4 with explanations in footnotes.

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*Canberra, A.C.T.*
1959 *October* 4

| Country | No. | Age | Rock unit | Sampling: Thickness in m. | Sampling: Lateral spread |
|---|---|---|---|---|---|
| Italy | 1 | Qr | Lavas Mt. Etna, 11 flows, 394 BC to 1911 AD (37·7 N, 15·0 E) | | |
| Sweden | 2 | Qr | Post-glacial varves Angerman river (63·1 N, 17·7 E) — 1 locality, 0 to 1000 AD*(1) | | |
| | 3 | Qr | Post-glacial varves Angerman river (63·1 N, 17·7 E) — 2 localities, 1100 BC to 750 AD*(2) | | |
| | 4 | Qp | Glacial varves — Central and S. Sweden, 8 localities, −50 to −3400 (de Geer)*(3) | | 850 km |
| | 5 | Qp | Glacial varves — Ragunda, 100 yr period about 9000 yrs ago*(1) | | |
| U.S.A. | 6 | Qp | New England varves, 11 localities, 15000 to 9000 BC (43 N, 73 W) | | 290 km |
| Japan | 7 | Qr | North Izu and Hakone Volcanic regions 35·5 N 138·5 E — Omuro-Yama | 700 | 400 km² |
| | 8 | Qp | Hakone — Central Cone | 700 | |
| | 9 | Qp | Hakone — Young somma | 300 | |
| | 10 | Qp | Hakone — Old somma | 700 | |
| | 11 | Qp | Yugawara | 300 | |
| | 12 | Qp | Taga | 500 | |
| | 13 | Qp | Usama — Upper | 300 | |
| | 14 | Qp | Usama — Middle | | |
| | 15 | Qp | Usama — Lower | | |
| | 16 | Qp | Kawajiri basalt, single eruption (34 N 131 E)*(1) | | |
| Iceland | 17 | Qr | Post-glacial lavas, 8 flows, ca. 5000 BC–0 | | |
| | 18 | Qp | Early Quaternary of W. Iceland, 9 flows, 3 localities, (64·6 N, 22·0 W)*(1) | | 10 km |
| | 19 | Tp–Qp | Plio-pleistocene Iceland (64·6 N, 22·0 W) — 33 normal flows | | |
| | 20 | Tp–Qp | Plio-pleistocene Iceland (64·6 N, 22·0 W) — 29 reversed flows | | |
| France | 21 | Tp–Qp | Auvergne lavas, 6 flows (45·1 N, 3·5 E) | | |
| Argentina | 22 | Q | Lavas of Neuquen, 10 flows (38 S, 70 W) | | 500 km |
| Australia | 23 | Tp–Qr | Newer Volcanics at Victoria, 32 sites (38·0 S, 143·5 E) | | 25000 km² |
| France | 24 | Tm–p | Lavas of Mt. Dore and Limagne, 5 flows (45·6 N, 3·0 E)*(1) | | |
| Iceland | 25 | Tm | Lavas of the N and NW, 102 flows at 5 localities (65·2 N, 20·0 W) | 700 | |

| No. Samples | No. Specimens | Stability | Treatment | Polarity + | Polarity − | Mean direction of magnetization $D$ | $I$ | $\alpha$ | Pole position Lat. N | Long. | Error $d\phi$, $d\chi$ | Reference |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 81 | | | nt | | 100 | 4·3 | +55·8 | 6·9 | 86·3 | 125·5 E | 7·3, 9·9 | 11, 38, 43 |
| | | | nt | | 100 | 357·4 | +73·4 | 3·3 | 86 | 142 W | 5, 6 | 5, 38 |
| ≃ 150 | | ac | nt | | 100 | 2·0 | +74·9 | 7·2 | 88·6 | 155·5 E | 12·0, 13·0 | 35 |
| 39 | | | nt | | 100 | | | | 84·3 | 141·1 W | 13·5, 13·5 | 34, 49 |
| 120 | | | nt | | 100 | 359 | +71 | | 82 | 160 W | | 35 |
| 1019 | | fd, ac | nt | | 100 | 355 | +51 | 10 | 81 | 132 E | 10, 14 | 50, 38, 45 |
| | | | | | 100 | 357 | +44 | 14 | 81 | 32 W | | 52 |
| | | | | | 100 | 351 | +55 | 23 | 86 | 65 E | | 52 |
| | | | | | 100 | 353 | +43 | 34 | 79 | 13 W | | 52 |
| | | | | | 100 | 349 | +53 | 11 | 83 | 50 E | | 52 |
| 300 | | ac | nt | | 100 | 357 | +46 | | 82 | 30 W | | 52 |
| | | | | | 100 | 341 | +55 | 20 | 78 | 62 E | | 52 |
| | | | | | 100 | 327 | +55 | 15 | 70 | 61 E | | 52 |
| | | | | 100 | | 146 | −43 | 29 | 64 | 35 E | | 52 |
| | | | | | 100 | 333 | +63 | 12 | 72 | 86 E | | 52 |
| 38 | | | nt | 81 | 17 | 167(S) | −56 | | 79 | 60 E | | 4 |
| ≃ 32 | | | nt | | 100 | 1·1 | +73·8 | 8·2 | 86·0 | 151·9 E | 13·3, 14·8 | 38, 43 |
| 51 | | | nt | 100 | | 181·0 | −75·2 | 6·9 | 87·0 | 149·0 E | 11·5, 12·6 | 38 |
| 33 | | | ac | | 100 | | | | 87·8 | 149·0 E | 6·7, 6·7 | 55 |
| 29 | | | ac | 100 | | | | | 77·3 | 73·9 E | 6·7, 6·7 | 55 |
| | | | nt | 100 | | 197 | −62·6 | 12·8 | 77·9 | 93·3 E | 15·7, 20·3 | 60 |
| 20 | 58 | ac | ac | 79 | 21 | 1 | −61 | 5 | 83 | 126 E | 6, 7 | 18 |
| 125 | 125 | | nt | 54 | 46 | 3·4 | −59·8 | 4·8 | 86·3 | 77·9 W | 5·5, 7·2 | 47 |
| ≃ 40 | | | nt | 100 | | 176·7 | −51·1 | 14·0 | 74 | 167 W | 13·6, 19·2 | 60, 45 |
| 102 | | | nt | ≃ 50 | ≃ 50 | 1·5 | +77·8 | 5·5 | 88·6 | 5·2 E | 9·6, 10·3 | 38, 43 |

| Country | No. | Age | Rock unit | Sampling: Thickness in m. | Sampling: Lateral spread |
|---|---|---|---|---|---|
| U.S.A. | 26 | Tm–p | Columbia River Basalts, 73 flows from 7 localities (47 N, 118 W) | 350 | 15000 km² |
| Canada | 27 | Tm–p | Lavas from the Yukon and British Columbia, 4 localities (61 N, 134 W)*(1) | | 500 km |
| New Zealand | 28 | Tp? | 7 Ignimbrite sheets at 3 localities in North Island (38 S, 176 E) | 140 | 13 km |
| Ireland | 29 | Te? | Lower Basalts of Antrim, 24 flows from 4 localities (55·1 N, 6·4 W) | >50 | 60 km |
| Scotland | 30 | Te? | Mull lavas, 8 flows (56·4 N, 5·8 W)*(1) | | 16 km |
| | 31 | Te–o | Mull intrusives, 5 units (56·4 N, 5·8 W)*(4) | | 12 km |
| | 32 | Te–o | Arran dykes and contacts, 5 dykes (55·5 N, 5·2 W)*(2) | | 10 km |
| England | 33 | T? | Lundy Island dykes and contacts, 5 dykes (51·2 N, 4·7 W)*(1) | | 4 km |
| France | 34 | To | Massif Central: Gergovie intrusives, 9 units (45·8 N, 3·1 E) | | |
| | 35 | To | Massif Central: Intrusive basalt of the Limagne, one intrusion (46 N, 3 E) | | |
| U.S.A. | 36 | Te | Siletz Volcanics, 8 flows (44·8 N, 123·7 E) | | 3000? km² |
| Australia | 37 | Te-o | Older Volcanics of Victoria (38·0 S, 145·5 E) | | 13000 km² |
| India | 38 | Ku-Te | Deccan Traps: Bombay State, 7 sites (18 N, 74 E) | | 300 km |
| | 39 | | Linga area, 9 sites in 4 flows near base of traps, (22·0 N, 78·9 E) | | ≃5 km |
| | 40 | | Khandala area, 20 sites in upper part of middle of traps (18·7 N, 73·4 E) | 630 | |
| | 41 | | Kambatki ghat, one flow (17·5 N, 74·0 E) | | |
| | 42 | | Igatpuri ghat, 3 flows (19·6 N, 73·6 E) | | |
| | 43 | | Nipani area, 2 sites in a lower flow (16·5 N, 74·3 E) | | |
| | 44 | | Nipani area, 5 sites in at least 3 upper flows (16·4 N, 74·4 E) | 50 | |
| | 45 | | Amba ghat, 5 sites in lower flows (17·0 N, 73·8 E) | 450 | |
| | 46 | | Amba ghat, 3 sites in upper flows (17·0 N, 73·8 E) | 10 | |
| | 47 | | Mt. Pavagadh, 8 lower flows (22·5 N, 71·6 E) | 100 | |
| | 48 | Tm? | Mt. Pavagadh, 2 sites in upper acid tuffs (22·5 N, 71·6 E) | 30 | |

| No. Samples | No. Specimens | Stability | Treatment | Polarity + | Polarity − | Mean direction of magnetization D | Mean direction of magnetization I | Mean direction of magnetization α | Pole position Lat. N | Pole position Long. | Pole position Error $d\psi$, $d\chi$ | Reference |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 138 | | | nt | 40 | 60 | 6 | +65 | 9 | 86 | 53 E | 4, 5 | 9, 19 |
| 46 | | | nt | 73 | 27 | 349 | +75 | 4 | 85 | 154 E | 5, 6 | 24 |
| ≈50 | | | nt | | 100 | 350 | −65 | | 79 | 38 E | | 37, 45 |
| 72 | | | nt | 100 | | 194 | −60 | 5·4 | 74 | 133 E | 6, 8 | 39, 19 |
| 63 | | | nt | 100 | | 166 | −73 | 18 | 82 | 69 W | 28, 31 | 8 |
| 50 | | | nt | 20 | 80 | 16 | +60 | 9 | 72 | 133 E | 11, 14 | 66 |
| 77 | | ct | nt | 71 | 29 | 5·5 | +63·2 | 9·4 | 78 | 156 E | 12, 15 | 51 |
| 21 | 42 | ct | nt | 100 | | 197 | −59 | | 74 | 123 E | | 7 |
| | | | nt | 100 | | 201 | −57 | 11 | 73 | 119 E | 12, 16 | 58, 38, 43 |
| | | | nt | 100 | | 180 | −73 | | 77 | 3 E | | 59, 45 |
| 57 | | fd, th | th | 63 | 37 | 70 | +55 | 7 | 37 | 50 W | 7, 10 | 16 |
| 45 | 45 | | nt | 33 | 67 | 17·0 | −72·9 | 6·8 | 66·8 | 57·3 W | 10·8, 12·1 | 47 |
| 7 | 14 | | nt | 71 | 29 | 149 | +56 | 10 | 28 | 78 W | 10, 15 | 43 |
| 69 | 171 | | nt | 100 | | 164 | +48 | | 37 | 83 W | | 13, 21 |
| 139 | 233 | | nt | 100 | | 147 | +50 | | 25 | 79 W | | 13, 21 |
| 4 | | | nt | 100 | | 176 | +60 | | 31 | 103 W | | 21, 45 |
| 12 | | | nt | 100 | | 161 | +51 | | 36 | 87 W | | 21, 45 |
| | 44 | | nt | 100 | | 168 | +60 | 4 | 35 | 95 W | | 21 |
| | 74 | | nt | | 100 | 338 | −32 | 5 | 52 | 70 W | 3, 5 | 21 |
| | 109 | | nt | 100 | | 144 | +60 | 7 | 27 | 76 W | 8, 10 | 21 |
| | 54 | | nt | | 100 | 335 | −26 | | 52 | 64 W | | 21 |
| 14 | 69 | | nt | | 100 | 351 | −16 | 8 | 58 | 91 W | 4, 8 | 21 |
| 8 | 15 | | nt | | 100 | 335 | +17 | | 75 | 89 W | | 21 |

| Country | No. | Age | Rock unit | Sampling: Thickness in m. | Sampling: Lateral spread |
|---|---|---|---|---|---|
| U.S.A. | 49 | Kl | Dakota ss, 1 site | | |
| Madagascar | 50 | Ku | Lavas and dykes, 10 sites | | ≃1000 km |
| Scotland | 51 | J | Lst. and ss. from Skye and Brora | | 200 km |
| Austria | 52 | Jm | North Calcareous Alps (47·6 N, 12·6 E) — radiolarite | 9 | |
| | 53 | Jl | North Calcareous Alps (47·6 N, 12·6 E) — red lst | 4 | |
| India | 54 | J | Rajmahal Traps, 3 quarries, Bihar (25·0 N, 87·9 E) | | |
| Rhodesia | 55 | J | Karroo basalt, 8 or 9 flows (18 S, 26 E) | | |
| South Africa | 56 | J? | Karroo dolerite — surface samples, 20 sills (30 S, 29 E) | | 30 000 km² |
| | 57 | | Karroo dolerite — Winkelhaak, upper sill (26·5 S, 29·1 E) | 23 | 1·5 km |
| | 58 | | Karroo dolerite — Winkelhaak, lower sill (26·5 S, 29·1 E) | 15 | 1·5 km |
| | 59 | | Karroo dolerite — Estcourt, reversed dolerite (29·1 S, 29·9 E) | | 2 km |
| | 60 | | Karroo dolerite — Estcourt, normal dolerite (29·1 S, 29·9 E) | | 2 km |
| Australia | 61 | J–K | Dolerite sills of Tasmania, 30 sites (42 S, 147 E) | >1000 | 20 000 km² |
| Uruguay | 62 | J? | Serra Geral lavas and baked Botacatu ss. (29 S, 57 W) | | 30 km |
| France | 63 | TRl | Vosge ss., 7 sites (48 N, 7 E) | | |
| England | 64 | TR | New Red Ss, 9 sites (53 N, 2 W) | | 400 km |
| Scotland | 65 | TR | New Red Ss., Arran (55·6 N, 5·3 W) | | |
| U.S.A. | 66 | TRu | Newark Series, Brunswickian Fm (41 N, 75 W) | | |
| | 67 | TRu | Lavas and sediments, Connecticut valley (42 N, 73 W) | | |
| | 68 | TRu | Lavas near Holyoke, Mass., 3 flows (42 N, 73 W) | | |
| | 69 | TRu | Springdale Ss. (37 N, 113 W) | | |
| | 70 | TRu | New Oxford Fm., Md., 1 site (39·5 N, 77·4 W) | 6 | 100 m |
| Bechuanaland | 71 | TRu | Kgoma Series (23 S, 27 E) | | |
| Australia | 72 | TRu | Brisbane tuff, 4 sites (27·5 S, 153·5 E) | | |

| No. Samples | No. Specimens | Stability | Treatment | Polarity + | Polarity − | Mean direction of magnetization $D$ | $I$ | $\alpha$ | Pole position Lat. N | Long. | Error $d\phi$, $d\chi$ | Reference |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 3 | 9 | | nt | 100 | | 164 | −62 | | 77 | 127 E | | 63 |
| 40 | | | nt | | 100 | | | | 68 | 168 W | | 62 |
| 6 | 14 | | nt | 67 | 33 | 226 | −66 | 6·6 | 56 | 76 E | | 54 |
| 15 | 21 | | nt | | 100 | 36·7 | +47·8 | 5·5 | 58·5 | 128 E | | 36 |
| 16 | 30 | | nt | | 100 | 47·9 | +50·6 | 6·5 | 53 | 112 E | | 36 |
| 33 | | | nt | | 100 | 328 | −64 | 5 | 13 | 70 W | 7, 8 | 15 |
| 12 | 44 | | nt | | 100 | 328 | −40 | 4·6 | 60 | 77 W | 3, 6 | 53, 20[7] |
| 33 | 149 | | nt | 30 | 70 | 352 | −62 | 12 | 76 | 128 W | 14, 19 | 33 |
| 17 | 57 | | nt | 100 | | 173 | +58 | 5 | 77 | 127 W | | 33 |
| 8 | 33 | | nt | | 100 | 307 | −63 | 12 | 44 | 98 W | | 33 |
| 15 | 53 | ct | nt | 100 | | 167 | +51 | 6 | 77 | 90 W | | 33 |
| 9 | 34 | ct | nt | | 100 | 331 | −64 | 8 | 67 | 86 E | | 33 |
| 60 | 60 | cg, ct | nt | [6] | 100 | 325 | −85 | 3·5 | 50 | 23 W | 7, 7 | 42, 1, 67 |
| 24 | 129 | ct, ac | ac | 28 | 72 | 354 | −43 | 3 | 83 | 126 W | 2, 3 | 18 |
| | 61 | | nt | 100 | | 218 | +9 | 12 | 28 | 143 E | 6, 12 | 14, 45 |
| 43 | 540 | ac, th | nt | 45 | 55 | 33 | +26 | 12 | 43 | 131 E | 7, 12 | 12, 51, 19 |
| | 41 | | nt | 100 | | 214 | −48 | 21 | 54 | 118 E | | 51, 45 |
| 21 | 71 | | nt | | 100 | 6 | +28 | 3 | 63 | 93 E | 3, 6 | 26 |
| 12 | 32 | | nt | 41 | 59 | 12 | +14 | 15 | 55 | 88 E | 8, 15 | 26 |
| 8 | 8 | | nt | | 110 | 10 | +14 | 11 | 54 | 90 E | 6, 11 | 26 |
| 8 | 18 | | nt | | 100 | 338 | +16 | | 55 | 107 E | | 63 |
| 14 | 14 | | nt | | 100 | | | | 62 | 158 E | | 32 |
| 8 | 44 | | nt | | 100 | 326 | −16 | 9·5 | 54 | 44 W | 5, 10 | 55 |
| 12 | 12 | ac | ac | 40 | 60 | 35 | −83 | | 39 | 37 W | | 48 |

| Country | No. | Age | Rock unit | Sampling: Thickness in m. | Sampling: Lateral spread |
|---|---|---|---|---|---|
| England | 73 | P | Exeter traps, 5 traps, Devonshire (51 N. 4 W) | | |
| Scotland | 74, 75, 76 | P | Ayrshire (55·4 N, 4·5 W): Mauchline lavas; Mauchline sediments; 5 kylitic intrusions*(1) | ≃100 | |
| Norway | 77 | P | Lavas of the Oslo graben (60 N, 10·5 E) | | |
| France | 78, 79, 80, 81 | P | Esterel (43·5 N, 6·8 E): pyromeride (R4); rhyolite (R3); dolerite; igneous and sedimentary rocks | | |
| | 82 | Pm? | Nideck porphyry, Vosge (48 N, 6 E) | | |
| | 83 | Pm | Montcenis Ss., Morvan (47 N, 4 E) | | |
| Germany | 84 | Pl | St. Wendel Ss., Sarre (47 N, 4 E) | | |
| U.S.A. | 85 | Pl-m | Yeso Fm., 1 locality, Tecolate, N. Mexico, (35·5 N, 105·3 W) | 10 | 60 m |
| | 86 | Pl-m | Culter Fm., Colo., 1 site (39·6 N, 107·4 W) | <1 | 15 m |
| | 87 | Pl | Abo Fm., Abo canyon, N. Mexico (34·5 N, 106·4 W) | 15 | 100 m |
| | 88 | Pl | Supai Fm.: Carizzo Creek (34·0 N, 110·3 W) | ≃300 | 16 km |
| | 89 | | Oak Creek Canyon (34·9 N, 111·7 W) | 30 | 2 km |
| | 90 | | Grand Canyon, Bright Angel and Kaibab trails (36·0 N, 113·7 W) | | 4 km |
| | 91 | | Grand Canyon, Kaibab trail and Flagstaff area (36 N, 113 W) | | 100 km |
| Kenya | 92 | Pu | Maji Ya Chumvi, Galana river (3 S, 39 E) | | |
| | 93 | Pl | Taru grit, Galana river and Taru quarry (3 S, 39 E) | | |
| Australia | 94 | Pm | Gerringong Volcanics, Upper Marine, Illawarra coast, 3 flows (34·7 S 150·8 E) | ≃200 | 15 km |
| | 95 | Pl | Lower Marine basalt flow, Hunter valley (32·7 S, 151·6 E) | 5 | |
| Britain | 96 | Cu | Clee Hills igneous rocks and baked sediments, 6 sites (52·5 N, 2 W) | | 50 km |
| | 97 | Cu | Tideswelldale baked sediments, Derbyshire (53·5 N, 1·5 W)*(1) | | |
| | 98 | Cu-Pl? | Lundy granite (51·2 N, 4·7 W) | | |
| | 99 | Cl | Kinghorn lavas, Fife (56 N, 3 W): 6 flows (41–46) | | |
| | 100 | | 5 flows (48–54) | | |
| | 101 | | 2 flows (64–65)*(1) | | |
| | 102 | Cl | Toadstones, Derbyshire, 3 volcanic units (53 N, 2 W) | | |
| | 103 | C | Sediments of the northern counties (53 N, 2 W) | | |
| | 104 | C | Shatterford intrusive, 2 sites (52·5 N, 2·5 W)*(2) | | 0·2 km |

| No. Samples | No. Specimens | Stability | Treatment | Polarity + | Polarity − | Mean direction of magnetization D | Mean direction of magnetization I | Mean direction of magnetization α | Pole position Lat. N | Pole position Long. | Pole position Error $d\psi$, $d\chi$ | Reference |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | 34 | | nt | 100 | | 189 | −9 | 20 | 43 | 164 E | 10, 20 | 17, 19 |
| 34 | | | nt | 100 | | 180 | −4 | 8 | 36 | 175 E | 4, 8 | 23 |
| 26 | | | nt | 100 | | 187 | −6 | 12 | 37 | 163 E | 6, 12 | 23 |
| | 8 | | nt | 100 | | 181 | +7 | | 31 | 174 E | | 2 |
| 12 | | ac | nt | 100 | | 202 | −33 | 9·6 | 46 | 165 E | 7, 11 | 64 |
| 5 | | th | nt | 100 | | 210 | −16 | | 46 | 141 E | | 61, 45 |
| 14 | | ac | nt | 100 | | 217 | −22·5 | 4·5 | 45 | 130·5 E | | 64 |
| 3 | | th | nt | 100 | | 175 | −13 | | 53 | 165 W | | 61, 45 |
| 14 | | th/ac | th/ac | 100 | | | | | 47 | 144 E | | 3 |
| 14 | 49 | | nt | 100 | | 193 | −7 | 4·5 | 43 | 168 E | 2, 5 | 56 |
| 3 | 14 | | nt | 100 | | 197 | +6 | 4·1 | 38 | 162 E | | 56 |
| 5 | 27 | | nt | 100 | | 181 | −9 | 3·4 | 45 | 175 W | | 56 |
| ≃20 | | | nt | 100 | | SE | low | | 41 | 127 E | | 32 |
| 2 | 10 | | nt | 100 | | | | | 36 | 121 E | | 32 |
| | 8 | | nt | 100 | | | | | 44 | 119 E | | 32 |
| 30 | 59 | | nt | 100 | | | | | 50 | 104 E | | 32 |
| | 22 | | nt | 100 | | | | | 42 | 122 E | | 32 |
| ≃20 | | | nt | 100 | | 146 | +8 | 7 | 39 | 115 E | 4, 7 | 22 |
| 20 | 34 | | nt | 100 | | 132·7 | +22·9 | 7·5 | 26 | 121 E | 6, 8 | 63 |
| 5 | 21 | | nt | 100 | | 267 | +38 | 11·3 | 4 | 150 E | | 20(?) |
| 8 | 32 | | nt | ? | ? | 87 | +61 | 16 | 0 | 87 E | 19, 25 | 20(?) |
| 19 | 19 | ac | ac | 100 | | 67 | +81 | 11 | 27 | 11 W | 21, 21 | 48 |
| 4 | 4 | ac | nt | 100 | | 110 | +80 | | 38 | 0 W | | 48 |
| ≃50 | | ct | nt | 100 | | 200 | +15 | | 27 | 155 E | | 14, 45 |
| 5 | | ct | nt | 100 | | 218 | +36 | | 9 | 143 E | | 14 |
| 5 | 10 | | nt | 100 | | 175 | −9 | | 43 | 180 E | | 7, 45 |
| 18 | | | nt | | 100 | 26 | −42 | | 8 | 153 E | | 14, 45 |
| 29 | | | nt | 100 | | 200 | +38 | | 11 | 158 E | | 14, 45 |
| | | | nt | | 100 | 20 | +15 | | 40 | 151 E | | 14 |
| 9 | | | nt | | 100 | 48 | +47 | 13 | 46 | 105 E | 11, 17 | 6, 45 |
| | | fd | nt | | 100 | 27 | +36 | | 51 | 143 E | | 6, 45 |
| 21 | | | nt | 50 | 50 | 33 | +3 | | 32 | 135 E | | 14 |

| Country | No. | Age | Rock unit | Sampling: Thickness in m. | Sampling: Lateral spread |
|---|---|---|---|---|---|
| U.S.A. | 105 | Cu | Naco Ss., Carizzo creek, Ariz | 60 | |
| | 106 | Cl | Barnett Fm., Texas (32 N, 99 E) — calcareous concretions at 8 sites | | 100 km |
| | 107 | | Barnett Fm., Texas (32 N, 99 E) — limestone, 1 site | | |
| Canada | 108 | Cl and u | Bonaventura, Kennebecasis and Bathurst Fms., of Gaspe and N.B. (48 N, 66 W)*(1) | | 350 km |
| | 109 | Cl | Codroy, red beds, Nfld. (48 N, 59 W) | | |
| Australia | 110 | Cu | Kuttung, Hunter Valley (33 S, 151 E) — varvoids from the glacial stage, 4 sites | >100 | 15 km |
| | 111 | | Kuttung, Hunter Valley (33 S, 151 E) — lavas from volcanic and glacial stages 3 flows, 5 sites | | 20 km |
| S. Rhodesia | 112 | Cu | Dwyka varves, Sebungawe 18 S, 29 E — normal | | |
| | 113 | | Dwyka varves, Sebungawe 18 S, 29 E — reversed | | |
| Britain | 114 | Dl and u | Old Red Ss. (52 N, 3 W) — 35 sites, S. Wales and the Border | >1000 | 200 km |
| | 115 | Dl | Old Red Ss. (52 N, 3 W) — 1 site in Brownstones | | |
| Australia | 116 | Su-D | Ainslie Volcanics, 4 sites A.C.T. (35 S, 149 E) | 40 | 2 km |
| U.S.A. | 117 | Sm | Rosehill Fm., 5 sites in Md. and W. Va. (40 N, 78 W) | | 40 km |
| | 118 | Sm | Clinton iron-ore, Ala, 1 site (33·6 N, 86·7 W) | | 100 m |
| China | 119 | Sm | Red sediments, Kansu | | |
| Australia | 120 | Su | Mugga porphyry, intrusive boss, A.C.T. (35 S, 149 E) | 20 | 100 m |
| Wales | 121 | Є1 | Caerbwby Ss., 10 sites (52N, 5 W) | 100 | |
| U.S.A. | 122 | Є(C) | Sawatch Fm., 2 sites, Colo. | | |
| | 123 | Єu | Wilberns Fm., 10 sites, Texas (31 N, 99 W) | | |
| Australia | 124 | Єm | Elder Mt. Ss., 1 site, W.A. (16 S, 128 E) | | |
| | 125 | Є1? | Antrim Plateau Basalts, 3 sites, W.A. (16 S, 128 E) | | |
| Britain | 126 | pЄ | Torridonian Ss. (58 N, 6 W) — Applecross and Aultbea groups, 81 sites | 3000 | 110 km |
| | 127 | | Torridonian Ss. (58 N, 6 W) — Diabaig group, 13 sites | 500 | 90 km |
| | 128 | pЄ | Longmyndian, 12 sites (53 N, 3 W) | | |

| No. Samples | No. Specimens | Stability | Treatment | Polarity + | Polarity − | Mean direction of magnetization D | Mean direction of magnetization I | Mean direction of magnetization α | Pole position Lat. N | Pole position Long. | Pole position Error $d\psi$, $d\chi$ | Reference |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 8 | 31 | | nt | 100 | | 149·7 | − 3·4 | 4·1 | 41 | 120 E | 4, 8 | 63 |
| ≅90 | | | nt | 100 | | | | | 41 | 128 E | | 40 |
| ≅8 | | | nt | | 100 | | | | 42 | 142 E | | 40 |
| 46 | 92 | | nt | 100 | | 163·5 | +19·5 | 5·0 | 30 | 133 E | 3, 5 | 25 |
| 9 | 36 | | nt | 100 | | 166 | +8 | 8·4 | 43 | 139 E | 5, 8 | 57 |
| 15 | 75 | fd | nt | 75 | 25 | 270 | −84 | 6 | 32 | 15 W | 12, 12 | 44 |
| 12 | 30 | | nt | | 100 | 5 | −85 | 8 | 43 | 30 W | 16, 16 | 48 |
| 4 | 10 | | nt | | 100 | 0 | −81 | 5·4 | 36 | 151 W | | 20(7) |
| 4 | 9 | | nt | 100 | | 333 | +76 | 6·9 | 7 | 17 E | | 20(7) |
| ≅100 | ≅1000 | fd | nt | 100 | | 196 | −5 | 5 | 30 | 159 E | 3, 5 | 17, 19 |
| 3 | 39 | | nt | 100 | | 234 | −22 | | 31 | 111 E | | 12, 45 |
| 12 | 12 | | nt | | 100 | 17 | −30 | 12 | 66 | 168 W | 7, 13 | 48 |
| 50 | | fd | nt | | 100 | 322 | −39 | 5 | 19 | 138 E | 4, 6 | 30, 19 |
| 7 | 16 | | nt | 100 | | SE | +ve | | 35 | 138 E | | 41 |
| 3 | 17 | | nt | | 100 | 293·6 | +55·3 | 8·5 | 49 | 12 E | | 10 |
| 9 | 9 | | nt | | 100 | 26 | −30 | 22 | 60 | 157 W | | 48 |
| 12 | 75 | | nt | 100 | | 187 | +39 | 8 | 15 | 173 E | 8, 10 | 17, 19 |
| 36 | | | nt | ≅100 | | SE | −ve | | 49 | 125 E | | 40 |
| ≅200 | | | nt | | | E | +ve | | 0 | 158 E | | 40 |
| 3 | 18 | | nt | 100 | | 231 | −15 | 10 | 34 | 164 W | | 48 |
| 7 | 28 | | nt | 66 | 33 | 53 | −2 | 12 | 36 | 154 W | 6, 12 | 48 |
| 205 | 1300 | cg, fd | nt | 58 | 30 | 303 | −44 | 5 | 6 | 43 E | 4, 6 | 46, 19 |
| 32 | 280 | cg | nt | | 100 | 307 | +34 | 7 | 35 | 118 W | 5, 8 | 46, 19 |
| ≅40 | ≅200 | ac | nt | 29 | 71 | 294 | −29 | 12 | 2 | 120 W | 7, 13 | 17, 19 |

| Country | No. | Age | Rock unit | Sampling: Thickness in m. | Sampling: Lateral spread |
|---|---|---|---|---|---|
| U.S.A. | 129 | pЄ | Baraga Co. dyke (3), Mich. | | |
| | 130 | pЄ-C | Keweenawan, Lake Superior, 6 groups — Upper — Chequamegon, ss. | | |
| | 131 | | Keweenawan, Lake Superior, 6 groups — Upper — Jacobsville, red ss. | | |
| | 132 | | Keweenawan, Lake Superior, 6 groups — Upper — Freda and Nonesuch, red ss. sh. | | |
| | 133 | | Keweenawan, Lake Superior, 6 groups — Upper — Copper Harbour, lavas and red ss. | | |
| | 134 | | Keweenawan, Lake Superior, 6 groups — Middle, Portage Lake, lavas | | |
| | 135 | pЄ | Hazel Fm., red beds, Texas — flat lying, 5 sites | | 3 km |
| | 136 | | Hazel Fm., red beds, Texas — folded, 9 sites | | |
| | 137 | pЄ | Hakatai Shales, Grand Canyon (36 N, 114 W) — Bright Angel trail | | 4 km |
| | 138 | | Hakatai Shales, Grand Canyon (36 N, 114 W) — Bright Angel and Kaibab trails | | |
| Canada | 139 | pЄ | Sediments, Nfld. (47 N, 53 W) — Signal Hill | | |
| | 140 | | Sediments, Nfld. (47 N, 53 W) — Blackhead | | |
| South Africa | 141 | pЄ | Pilansberg dykes, 5 dykes, $1290 \pm 180 \times 10^6$ yrs. | | |
| | 142 | pЄ | Bushveld gabbro, 5 quarries, $2 \times 10^9$ yrs. | | 200 km |
| Australia | 143 | pЄ | Buldiva Quartzite, N.T. (14 S, 132 E) | | |
| | 144 | pЄ | Nullagine lavas, W.A. (21 S, 120 E) | | |
| | 145 | pЄ | Edith River Volcanics, 10 sites, N.T. (13 S, 132 E) | | |

| No. Samples | No. Specimens | Stability | Treatment | Polarity + | Polarity − | Mean direction of magnetization $D$ | $I$ | $\alpha$ | Pole position Lat. N | Long. | Error $d\psi$, $d\chi$ | Referent |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| ≃50 | | ac | nt | 100 | | 82 | −86 | 1 | 45 | 99 W | | 31, 19 |
| 15 | | | nt | | 100 | 30 | +74 | 6 | 69 | 47 W | 10, 11 | 23 |
| 15 | | | nt | | 100 | 250 | −11 | 13 | 14 | 10 E | 7, 13 | 23 |
| 68 | | fd | nt | | 100 | 285 | −1 | 3 | 9 | 169 E | 2, 3 | 23 |
| 25 | | cg | nt | | 100 | 294 | +32 | 7 | 30 | 176 E | 5, 8 | 23 |
| 31 | | | nt | | 100 | 282 | +41 | 4 | 25 | 170 W | 3, 5 | 23 |
| 15 | | | nt | | 100 | NW | +ve | | 49 | 175 W | | 41 |
| 39 | | | nt | | 100 | | | | 60 | 151 E | | 41 |
| 15 | 34 | | nt | | 100 | 267·7 | +72·8 | 5·4 | 31 | 150 W | 9, 10 | 63 |
| | | | nt | | 100 | 246 | +72 | 17·5 | 18 | 144 W | 27, 30 | 22 |
| 9 | 36 | | nt | | 100 | 283 | +20 | 11·3 | 16 | 142 W | 6, 12 | 57 |
| 10 | 40 | | nt | | 100 | 232 | +51 | 9·9 | 5 | 84 E | 9, 13 | 57 |
| 169 | | ac | nt | | 100 | 24·0 | +69·3 | | 7·5 | 42·5 E | | 28 |
| 99 | ≃200 | fd | nt | | 100 | | | | 23 | 36 E | | 29 |
| 8 | 8 | | nt | 100 | | 243 | +38 | 12 | 30 | 121 W | | 48 |
| 5 | 15 | | nt | 100 | | 143 | +64 | 8 | 51 | 18 W | | 48 |
| 10 | 10 | | nt | 100 | | 90 | +48 | 18 | 6 | 14 E | 15, 24 | 48 |

(1) Pole calculated from the mean directions given in the originals cited.
(2) Mean direction and pole calculated from data given in the originals cited.
(3) Pole calculated as a mean of poles obtained from the directions at 8 localities given by Granar (34), page 27.
(4) Pole calculated from the mean direction of 3 intrusives (olivine gabbro, granophyre and felsite ring-dykes) given by Vincenz (66) table 1.
(5) Reversed specimens only.
(6) A few reversals recorded.
(7) Nairn's data.

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# On Some Movements of the Ground in Geneva

A. Decae*

*Summary*

This article reviews the methods adopted for measuring ground motions and the results obtained during the construction and alignment of the 25 GeV proton synchroton at CERN.

The 25 000 million electron volt synchroton (the Eurotron) that has been under construction by the European Centre for Nuclear Research (CERN) in Geneva these last five years could not be a mere scale-up of the existing machines of smaller energy. The cost of the magnet would have been prohibitive. A new method of accelerating particles was found in 1954 which allowed a considerable reduction of the size of magnets and coils necessary to keep the particles on a circular orbit. This is the so-called method of alternating gradients, using successive focusing and defocusing magnetic units. It is well known that series of convergent and divergent lenses have an overall converging effect, and so is the case with magnetic units, the net advantage being an important reduction of the excursions of the particles about their mean orbit, and consequently the reduction of the dimensions of the magnets and coils. But, and this is the drawback, such an arrangement demands an accurate alignment of the magnets on their circle. In fact, to limit the excursions of the particles due to misalignments to 1 cm—and as there are many causes for the protons to wander off their orbit, it is rather necessary to limit the maximum effect of each one to $\pm$ 1 cm if the protons are to be kept inside the vacuum chamber—it is necessary to set up the magnets within $\pm$ 0·1 mm of their true position. Now as they are 40-ton blocks, 4 m long, disposed on a 100 m radius circle, this calls for certain precautions.

The first step was obviously to study the stability of the superficial layers by means of a geodetic net stretched between solid monuments firmly imbedded 1 m deep to escape frost and thaw deformations.

The ground at the chosen site—chosen for political, economical, financial, and other reasons, for all but geological ones—consists of Quaternary moraines, 4 to 15 m deep, resting upon the Tertiary molasse (oligocene chattien), deeply indented with ridges and gulleys. The molasse becomes progressively sandstone as one gets deeper until the chalk is reached at 500 m down, and it fills the whole bed of the Würmst Rhône glacier between Jura and Salève. Several anticlinals of molasse lie within this trough, oriented as the present Rhône valley and Jura, Salève ridges, SW–NE. CERN site lies on the rim of such an anticlinal, the Choully hill, where the moraine is particularly thin, as little as 1 m in some places.

One word must be said also of the methods and instruments used to reach the expected accuracy. Since the magnets were to be set at $\pm$ 0·1 mm, it was necessary

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to perform all elementary operations to a few hundredths of a millimetre accuracy, and to secure the stability of the foundations within the same limits. To give a sense to this accuracy in ground measurements, it was imperative to determine particular points of the ground within 0·01 mm. To do this, bronze plates were inserted in the concrete at the top of monuments, and cylindrical cavities carefully ground into them. On the other hand, all geodetic instruments, theodolites,

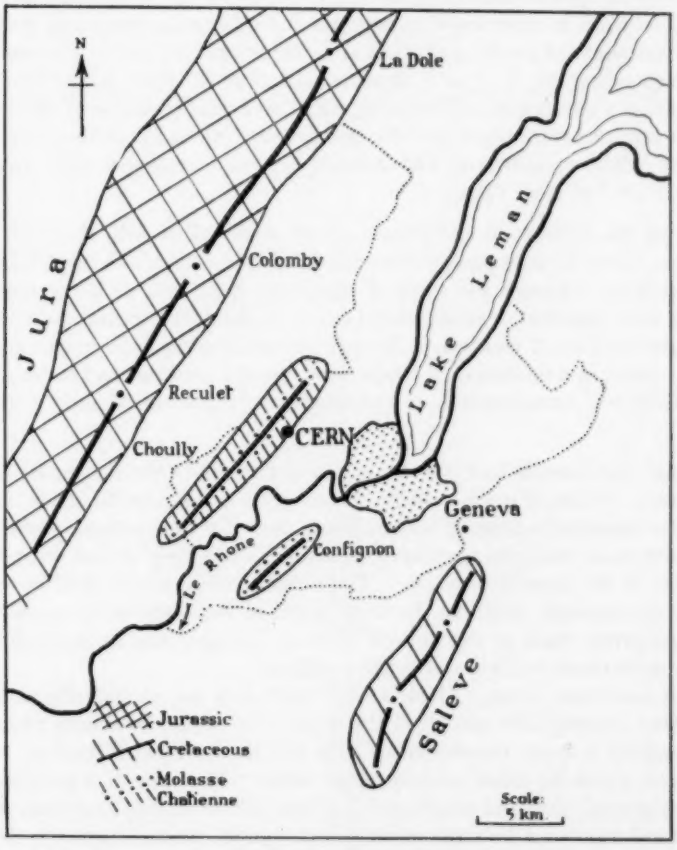


FIG. 1.—Geneva anticlinals.

targets, microscopes were fitted with small balls at their basis, exactly centred with their axis, which could be forced into the cavities, thus securing the permanence of the point of station. Steel balls were used at first, of exact fitting, but in time backlash occurred that marred the required accuracy, so nylon balls were substituted, with a diameter larger by a few microns, thus securing a stable and perfect centring (within a few microns).

Various improvements were also devised to the usual instruments, to speed up the measurements, such as:

fitting T-3 theodolites with small very well centered targets to allow reciprocal bearings;

combining in a single instrument a theodolite and a cranked microscope, so that readings could be taken simultaneously on angles and lengths on the lines engraved at the end of invar wires;

using as targets carefully centered collimators focused at a standard distance so as not to bring about any change of focus during angle measurements;

developing special pulleys to stretch invar wires: good ordinary ball-bearing pulleys, used at metrological institutes for calibration purposes, bring about an uncertainty of about $\pm 10$ g in a 10 kg force tension, due to internal friction, corresponding to a $\pm 10^{-3}$ random variation of their elastic elongation—35 $\mu$m for a 100 m wire. This being much too large for our requirements, knife-edge supports were fitted into the pulleys' axes, upon which they were made to swing when equilibrium was reached, so that tensions were transmitted within $10^{-6}$ of their value;

making the calibration and measurement observations with the invar wires at known times so that plastic elongations of this rather soft material could be eliminated: although the limit of elasticity ought not to be reached before 60 kg force tension on usual wires (2 mm$^2$ section), the special pulleys allowed the observations of continuous elongations under 10 kg force tensions, reaching a few microns a minute on a 100 m wire. So the readings had to be scheduled carefully on measurements and calibrations to get rid of such a systematic drift.

Special instruments had also to be devised to study the deformations of the deep layers. When it grew evident the magnets could not be made simply to rest on the moraine because of its instability, and that foundations would have to reach down to the underlying molasse, it became interesting to find out how stable these were in the three coordinates. The obvious thing was to drill borings near geodetic monuments, down to the 20 m depth of foundations, to establish some sort of reference mark at the bottom of these borings, and measure differential movements between surface and depth markers.

Easier said than done. These CERN moraines are abundantly aquiferous, and molasse disintegrates quickly in the presence of water, becoming clammy and fluent, making it quite impossible to keep dry the bottom of borings, whatever precautions could be taken to keep water away. The idea of a pendulum suspended at ground level and reaching down 20 m where its displacements would be detected and measured by some means (electrostatic, optical, magnetic, electronic, etc.) had to be given up.

So making the best out of a bad job, this unavoidable water was used to float at ground level hollow spheres of copper (of 40 cm diameter) stretching upwards invar wires fastened with araldite to the bottoms of borings—incidentally, these borings had to be straight and vertical, which is not very easy. These spheres follow the movements of their moorings at the bottom, and crosshairs, fitted on the top of them, are observed through sliding vertical microscopes, fixed to the superficial layers, that allow the measurement of relative displacements with a 0·01 mm accuracy.

The range of focus of the microscopes was small enough to allow them to be used to measure the changes in elevation.

The geodetic observations at ground level have revealed the existence of important deformations. The geodetic monuments—7 in all—covered an area

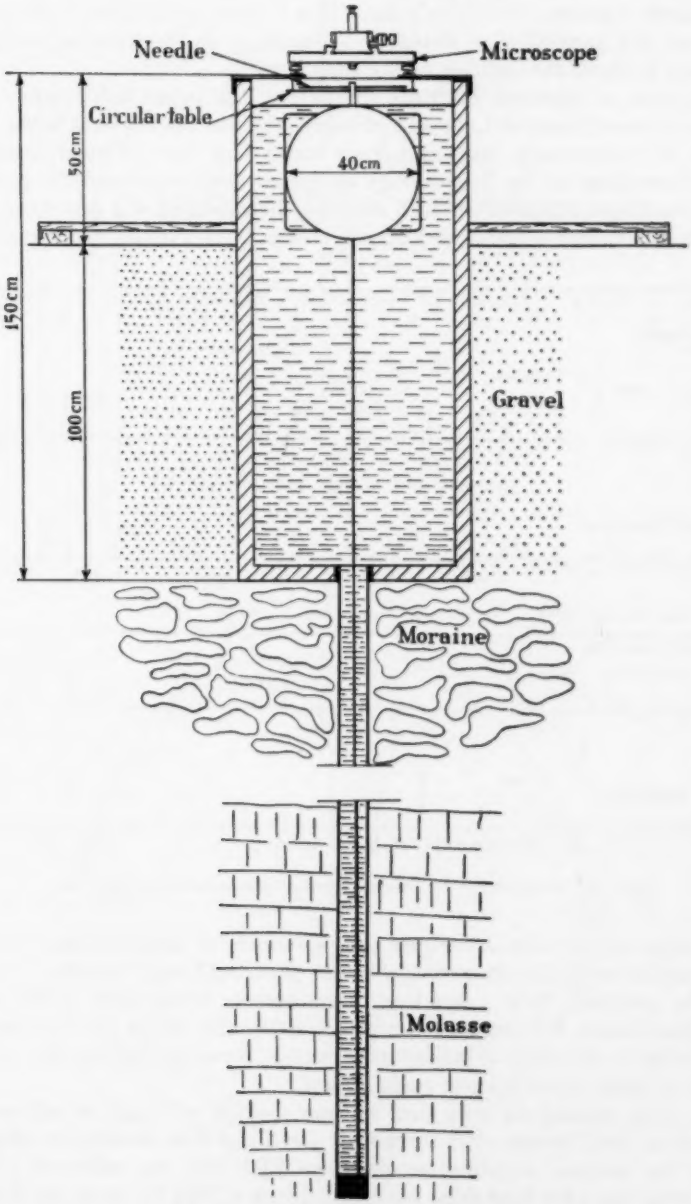


FIG. 2.—Sketch of a depth pendulum.

about $\frac{1}{4} \times \frac{1}{8}$ mile around the proposed site of the synchrotron and the changes of length reached sometimes as much as 2·5 mm over the whole length, i.e. a little more above 1/200 000 in relative value. This is quite considerable and obviously ruled out any possibility of founding the machine on the moraine, rendering it necessary to check the stability of the deep layers.

Moreover, it appeared from the results that the larger deformations of the moraine followed more or less spells of heavy rainfalls. As the local folklore is full of tales of subterranean lakes and rivers and a deep sheet of water, even to the point of ascribing to the Jura springs an Alpine origin—pretendedly proved by their mineralogic composition—and asserting the existence of a subterranean lake beneath Lac Leman connecting Jura and Alpine waters under high pressures, it

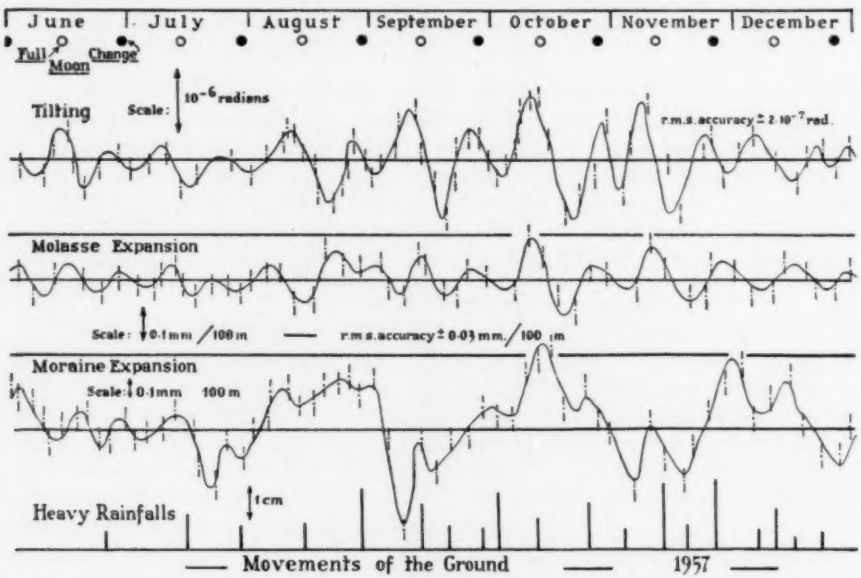


FIG. 3.—Correlation of different ground movements on the site.

was thought worth it to investigate the possibility of deformations of the hill containing the site under the action of waters that could swell or deflate it according to its pressure. It is a fact that in springtime, when snow melted on the nearby Jura slopes, deformations increased considerably in size giving appearance of credibility to the story of subterranean waters, however unlikely the existence of sheets of water inside molasse could appear.

The depth pendulums were then set into position to check on subterranean deformations and, besides that, systematic levellings were undertaken along the ridge of the molasse anticlinal supporting CERN site, the *colline de Choully*. Once a week, this 4 km long ridge was levelled with a Wild N3 level, the distances between bench marks being kept very short (40 m) in order to reach a very good accuracy (rms 0·7 mm/(km)$^{1/2}$). All benchmarks had thus their elevation constantly checked during the best part of two years, providing the outline of a vertical cut of the hill. It was rather expected to find some sort of swellings, after rains

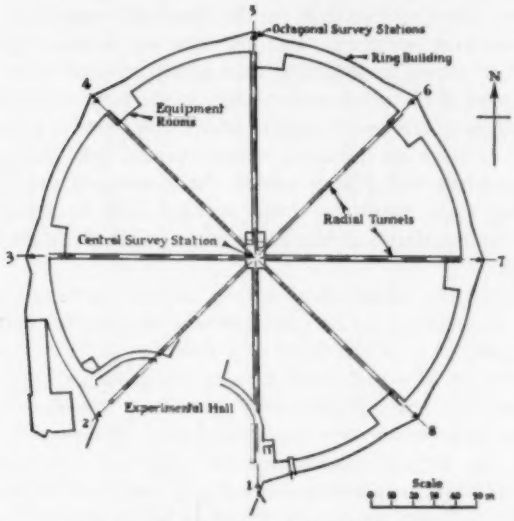


FIG. 4.—General plan of P.S. building.

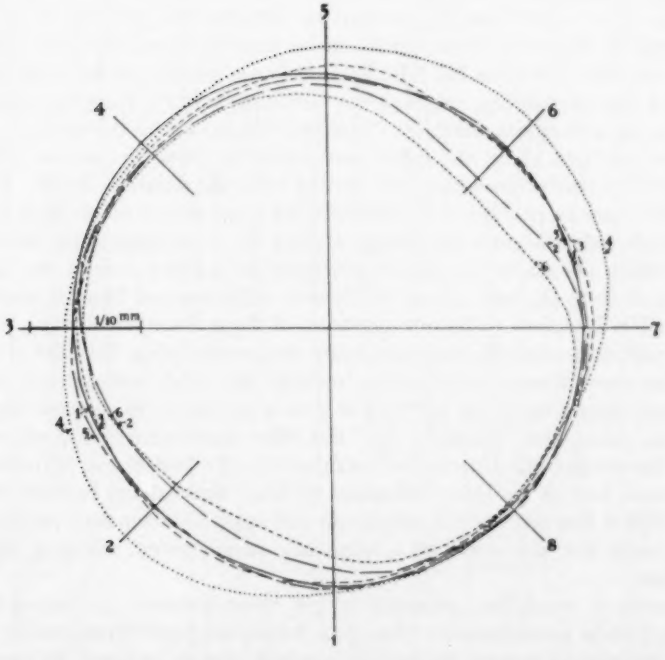


FIG. 5.—Periodic displacements of octogonal monuments due to ground expansions and contractions. The maximum deformations are felt along a 2–6 direction, parallel to the general direction of the Rhône talweg, while they are practically non-existent in an orthogonal direction.

or melting of snow, that could account for the observed elongations of the moraine. Nothing of the sort ever occurred. The hill retained its exact shape all the time. But it was found to swing periodically, as a whole, around its ends in a vertical plane. The changes of elevation were always in the same ratio as the horizontal distances, reaching an overall amplitude of about 7 mm over a 4 km distance. The rains were found to have no influence whatsoever on this tilting, the period of which was approximately half a lunar month. An increase of amplitude was indeed observed at spring time, when snow was melting, and in autumn, when rains were heavy, but the regularity of the phenomena made it rather hard to ascribe it to a local cause.

On the other hand, the subterranean survey carried out with depth pendulums showed that the deep layers underwent periodic expansions and shrinkings of about 0·5 mm amplitude over a quarter of a mile ($\sim 10^{-6}$) with a period which was also approximately of a half lunar month, which rather suggested that both phenomena—expansion and tilting—were somehow connected. No vertical movements of the deep layers were ever found out. Horizontally, the phase lag is about a week, the molasse undergoing its maximum deformation when the tilting is at its mean value, expanding when the ground is tilting down eastwards, shrinking when it tips down westwards, which is rather hard to understand since one should have rather expected symmetry in this matter and a period of expansion half that of tilting. It is true the expansion measurements are rather less reliable than the tiltings, since the amplitude of observed phenomena is nearer to the sensitivity of the method and the accuracy that can be expected of it, but numerous checkings were carried out that seemed to confirm this paradox of asymmetric comportment. Moreover, these deformations were found anisotropic, as well, and to occur only in a direction NE SW, parallel to the general direction of the Rhône valley and the surrounding ridges of the Jura and Salève. They were quite non-existent in an orthogonal direction. This was figured out as due to some stretching of the molasse along the gully constituted by these cretaceous walls when some sort of perturbation entered it, coming from the Atlantic Ocean. For these movements were ascribed to indirect effect, for want of any other likely cause.

To check independently the tilting, we had the recordings of the level of Lac Leman, which are taken at a dozen limnigraphic stations around the lake, forwarded to us from the two stations of Geneva-Sécheron and Morges, some 40 km away in a NE direction. The interpretation of these documents over a period of years proved rather complicated since many causes may bring the level of the lake to vary between Morges and Geneva, namely the wind, atmospheric pressure, the current, depending on the aperture of sluices at Geneva to regulate the Rhône output, the tides, the "seiches", etc. But after accumulating records of these different factors and eliminating their influence to the best of our knowledge and ability, some sort of periodic difference of level showed up between the two stations, with a few centimetres amplitude and again a half month period, which seems to mean that this could be a large-scale phenomenon, affecting the whole tectonic area.

Of course, it would be necessary to get more accuracy, to use specialized apparatus, such as extensometers, tiltmeters, horizontal pendulums, gravity meters, etc., and the present results, reached by geodetic means, can only be considered as indications; but, anyway, they were sufficient for the purpose we had in view, namely the foundations of the 25 GeV synchrotron and the alignment of its magnets within 0·1 mm of their specified position.

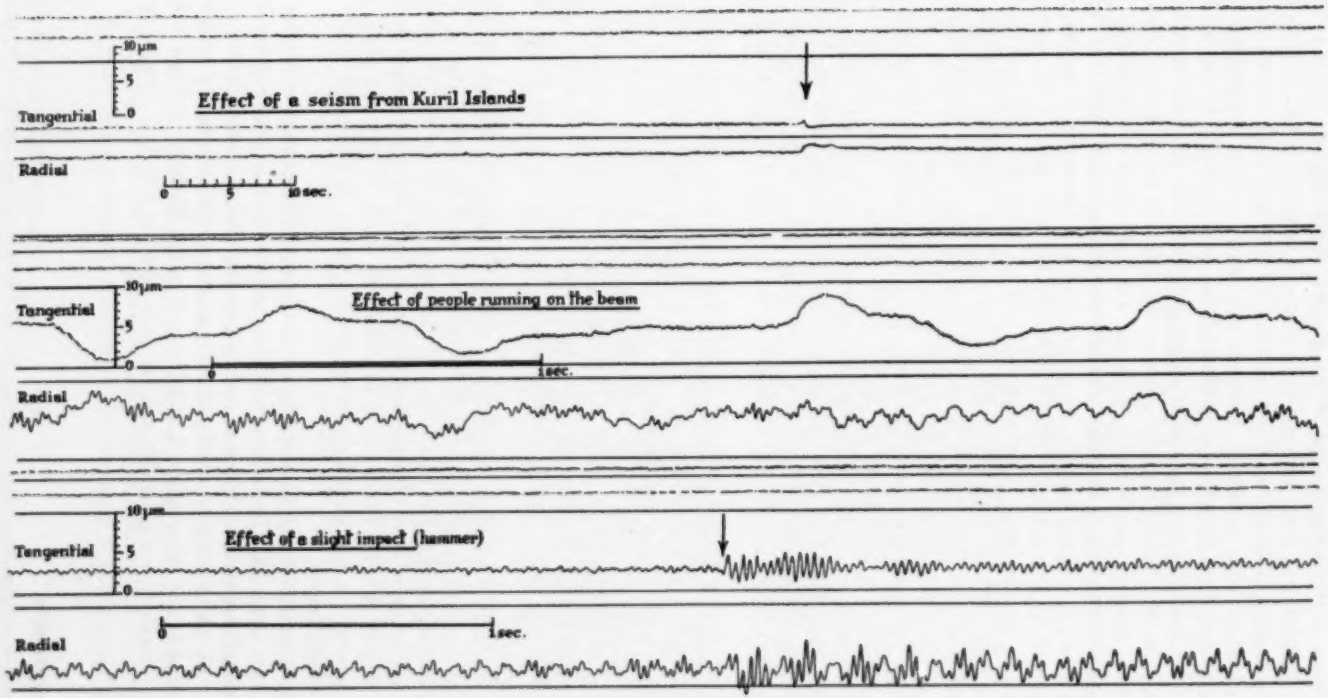


FIG. 6.—Some short-period movements of the beam recorded with inductive capacitors.

As, in view of the deformations of the ground, these magnets could not have individual foundations, they were set on a circular concrete beam, resting itself on 80 piles reaching to the molasse 15 m down. To free it of inopportune stresses, elastic suspensions were provided, between this beam and its supporting piles, allowing it to retain its circular shape through its own rigidity. It was enclosed in a tunnel for shielding and air-conditioning purposes, and geodetic monuments were erected in this tunnel founded also on deep layers by means of concrete pillars. These geodetic monuments, 8 in all, are assured of intervisibility with their immediate neighbours through the circular tunnel, and with the centre monument through 8 radial tunnels, thus constituting 8 first-order triangles, with angles measured by theodolites Wild $T_3$ and sides by invar wires with graduated reglettes under microscopes. The magnets, laid on oil jacks, and provided on their upper surfaces with bronze sockets related to their magnetic elements, are then aligned from the monuments inside the ring by angle and length measurements, all operations being conducted within 0·02 mm accuracy.

Of course, the deformations of the ground must be taken into consideration, since the monuments participate in them. Their distances from the centre are checked before each survey and the necessary corrections are applied. The changes of radius are thus found to approximate 0·1 mm over 100 m, and the concrete beam follows these movements through the restoring forces of its elastic supports within 0·02 mm of their amplitude. Corrections are then easy to apply to measurements.

Another variety of ground movements caused some worry when first recorded, namely small seisms and microseisms. About a hundred small seisms a year are felt at Geneva with a 0·01 mm amplitude, half of them being from the nearby Valais with somewhat sharp fronts and noticeable accelerations. But they last only a matter of minutes and through its inertia the concrete beam remains usually undisturbed (the magnet units being only laid on their jacks, their stability resting on their weight (40 tons), stiff horizontal accelerations in the beam could easily dislodge them).

But microseisms were more troublesome and particularly a continental agitation (amplitude 10 $\mu$m, period 4 s) which sometimes lasted for days running, even a week. This was found out to be induced by the pounding of the surf on Atlantic beaches whenever gales were blowing from the west, and our vibrometers showed that after a day or two, the beam began to vibrate by itself, in as much as its own period lies around 1 s. So special damping devices had to be rigged up to get rid of this new disturbance.

It can be seen that the ground at Meyrin is the worst that could be chosen for such a project. Even the Brookhaven 25 GeV synchrotron is in better position with its homogeneous glacial Moraine soil and sands. But the best ground of all would be chalk, good compact chalk, with its mixture of hardness and resilience not to be found in any other ground, and Nimrod, at the Rutherford High Energy Laboratory, Harwell, lies probably on the best possible site ever chosen for a synchrotron.

**Acknowledgments**

I am greatly indebted to Mr Gervaise, from Institut Géographique National, for his very accurate measurements, and to Mr Augsburger, chief of CERN workshop, for his keen cooperation in mechanical fittings and gadgets.

# Reports of Meetings

## Meeting of the International Gravity Commission 1959

### 1959 September 15–19

The International Gravity Commission held its third meeting in Paris, at the International Gravity Bureau from the 15th to the 19th September 1959.

In 1948, at the General Assembly in Oslo, the Section of Gravity of the International Association of Geodesy (Chairman, Professor Vening Meinesz and Secretary, Father P. Lejay) showed interest in collecting the gravity data which were getting more and more important as apparatus was technically developed and the measurements became easier. It expressed the hope of the creation of an International Gravity Bureau and of an International Gravity Commission, the purposes of which would be the selection and the collecting of all gravity data available in each country.

In 1951, at the General Assembly in Brussels, this resolution was realized and in 1953, under the Chairmanship of Father Lejay, Director of the International Gravity Bureau and President of the Section of Gravity, the International Gravity Commission held its first meeting in Paris and laid the foundations of a homogeneous world net of gravity measurements. Since this first meeting, gravity has been considerably developed and owing to the various problems examined during the third meeting it seems difficult to report everything in full detail.

The symposium was opened by Professor Cassinis, President of the International Association of Geodesy, Professor P. Tardi, Director of the Central Bureau of the International Association of Geodesy and of the International Gravity Bureau, and Mr B. C. Browne, President of Section IV (Gravity). Twenty-five nations were represented. Apart from the delegates of the International Gravity Commission (one member for each nation) many interested scientists attended the meeting.

Three main purposes dominated the organization of the meeting:

The necessity of getting the gravity world net with exactitude (absolute measurements), with accuracy and homogeneity (techniques of measurements by means of pendulums and gravity meters, calibration bases for gravity meters and accurate connexions between the main stations);

the extension to unexplored areas and especially to the ocean surfaces;

the use of the gravity data obtained from the observation of artificial satellites in order to get a better knowledge of the problems of geodesy and geophysics.

#### Absolute measurements

*Dr A. H. Cook* presented his report and summarized the absolute measurements still in progress. He stressed the fact that the difficulties of absolute measurements do not lie in the techniques of length and time measurements as such (in

modern work atomic standards of length and time are employed) but in the definition of the quantities to be measured and in the elimination of forces other than gravity. With the reversible pendulum the relevant times and distances are very well defined but systematic errors due to elastic deformation of the support and of the pendulum itself, and, most serious, the imperfectly understood forces acting at a knife-edge support, are liable to occur. In the method using the free fall of a rod, the definition of the events timed is rather uncertain and the geometrical adjustment of the optical system is very critical. The effects of residual air in the enclosure and the motion of the apparatus due to ground movements are much more serious than for pendulum observations. The symmetric (or "return ticket") free motion method is better than the simple fall method in all respects.

The following list gives for each absolute determination the correction to be added to the reference value adopted at Potsdam:

| *Authors* | *Site* | *Method* | *Result* |
|---|---|---|---|
| Heyl & Cook 1936 | Washington | Reversible pendulum | −16·8 mgal |
| Clark 1939 | Teddington | Reversible pendulum | −12·8 |
| Agaletskii & Egorov 1956 | Leningrad | Reversible pendulum | −12·1 |
| | Leningrad | Free and constrained fall | −9·3 |
| Martsinyak 1956 | Leningrad | Free fall | −8·4 |
| Volet 1952 | Paris-Sèvres | Free fall | −24·8• |
| Thulin 1958 | Paris-Sèvres | Free fall | −12·8 |

## Techniques of relative gravity measurement apparatus

Reports on techniques of pendulum apparatus and gravity meters were made by *Mr B. C. Browne* and *Mr J. Martin* respectively.

It is worth while noticing the new tripendular apparatus of the Italian Geodetic Commission of which the pendulums are molybdenum.

Among the results reported, information from *Dr Woollard* concerning the LaCoste Romberg gravity meter should be mentioned (slow drift: about 0·5 mgal per month) as well as the systematic studies on the variations of calibration of gravity meters with temperature and time: for instance for the Worden gravity meter a variation of 1 in $10^3$ for 10° C temperature variation and an annual variation of 1/2000 have been recorded.

In conclusion *Mr J. Martin* said that the average limiting accuracy at present obtainable with gravity meters was 0·05 mgal and he expressed the hope that the results of international interest obtained with gravity meters would be published together with all information necessary to give an idea of the actual accuracy of the measurements.

## Calibration bases

*Professor M. Kneissl* presented a complete report on the European Calibration Base for gravity meters of which the main line stretches from Bødo to Catania and the secondary line from Teddington through Paris to Pic du Midi.

*Dr Marzahn* made a complementary report on the methodical study of measurements by means of pendulums and gravity meters. He said in conclusion that the

• Provisional result.

accuracy of the adjusted gravity differences between the adjacent pendulum stations is about $\pm 0{\cdot}3$ mgal, and that a root mean square error of the calibration factors of $3 \times 10^{-4}$ may be expected; so on the gravity difference Bødo–Catania (about 2 300 mgal) a root mean square error of $\pm 0{\cdot}7$ mgal may occur, essentially determined by the uncertainty of the calibration factor.

*Dr D. A. Rice* made a brief review of the calibration bases used in North America (fundamental base from Mexico to Point Barrow in Alaska and practical bases of Canada and U.S.A.). The comparison between these practical bases was achieved in 1956–57 by the Coast and Geodetic Survey and it has been concluded that the U.S. calibration system was about 5 parts in $10^4$ higher than the Canadian calibration system.

In conclusion it was recommended that all calibration bases should be completed by new pendulum measurements using the various types of pendulums.

## First order world net

*Professor C. Morelli* presented a very detailed report on "Absolute and First Order World Net".

Following the discussions, various resolutions were proposed and adopted: first that the height corresponding to the conventional Potsdam value 981·274 0 gal be conventionally understood to be 87·00 m and that the value of gravity at Bad Harzburg, auxiliary station of Potsdam, be, for the time being, conventionally adopted as 981·274 0 gal; secondly that only those connexions with an accuracy of better than 0·5 mgal should be taken into account in the establishment of the world gravity net and that the auxiliary stations should be connected to the main station with an accuracy of 0·03 mgal; lastly, that the following pendulum interconnexions should be carried out: Buenos Aires to Mowbray, and to Christchurch or Melbourne, Helsinki to Hammerfest and Potsdam, Rome to Beirut and Singapore; and that the connexion previously proposed between Helsinki and Leningrad should be carried out by means of gravity meters.

## Gravity measurements at sea

*Dr J. Lamar Worzel* gave a general report on the recent gravity measurements at sea.

He reviewed gravity measurements in shallow areas with gravity meters supported on the ocean floor (measurements in the Bahamas, in the Baltic and Barents Seas, in the Adriatic, Ionian and Tyrrhenian Seas and in the Moray Firth), then submarine gravity measurements (measurements between Australia, New Zealand across the Tonga Trench to the Fiji Islands and back to Sidney, measurements in the vicinity of Bermuda, measurements near the coasts of Central America, measurements in North Sea and in Western Mediterranean and in Eastern North Atlantic).

New information was given on gravity measurements on surface ships. Particularly *Professor Tsuboi* reported on improvements concerning the surface ship gravity measurements made in Tokyo Bay by *Mr T. Tomuda* and himself.* "For stabilizing the gimbals to which the pendulum case is fixed with a required accuracy, we are now using inertial rotors which are mounted coaxially with the gimbals and

* See *Geophys. J.*, 1 (1958), pp. 97–98.

the rotational speeds of which are controlled by a servomechanism driven by horizontal accelerometers." It is hoped that a test voyage may be carried out late in 1959 or early in 1960.

On the other hand in 1957 the Lamont Group using a Graf sea gravity meter mounted on a gyro-stabilized platform obtained the first observations on a surface ship in the open sea. Since this initial work, much additional material has been obtained and *Dr Worzel* reported it in detail and showed slides to demonstrate the comparison between the observations with the Graf sea gravity meter on USS *Compass Island* and Vening Meinesz pendulum stations previously made on a submarine. USS *Compass Island* is an 8 000 ton fin-stabilized naval auxiliary. Nine runs as nearly as possible along the same line of pendulum stations south-east of New York showed a total spread of 20 mgal. Two meters operating on the same stable platform showed only 1 mgal difference as a maximum. One half of the spread is attributed to navigational problems such as the calculation of the Eötvos correction, and the other half to the geographic locations of the runs. The soundings also showed a spread. Observations over such features as sea mounts, Mid-Atlantic ridge, Canary Basin and through the Straits of Gibraltar were shown. Finally slides of a research project still in progress concerning world-wide gravity anomalies at sea were also presented.

*Dr Worzel* also reported that 6 000 miles of continuous gravity records were obtained with a Graf Sea gravity meter installed on a jury-rigged stable platform, and that twenty-two surface ship gravity measurements were made with the LaCoste Romberg submarine gravity meter and that other measurements were made with a LaCoste Romberg surface ship meter along a 300 mile track off southern California (by J. C. Harrison).

*Dr Graf* himself gave information concerning his apparatus and presented his report "Über die Möglichkeit der Kompensation der Horizontalbeschleunigung auf Schiffen bei Seegravimetermessungen".

Concerning measurements in the bathyscaphe, *Professor Morelli* recalled the observations carried out in the Gulf of Messina and *Mr J. Martin* said that he made a test measurement at 2 400 m depth in the French bathyscaphe without any difficulty in the Mediterranean Sea.

## Airborne gravity measurements

*Dr Thompson* described the test flights of airborne gravity measurements carried out in order to discover if the accuracy required for geodetic purposes can be reached. The experiment was conducted at Edwards Air Force Base, California (1958 November 6–7) with a LaCoste Romberg sea gravity meter (surface type). The meter itself was suspended from a gimbal joint and therefore always hung in the resultant vertical when a horizontal acceleration was acting. The apparatus was mounted on the floor of a KC-135 jet tanker very near the centre of gravity of the aircraft. Flight control data were provided by a Doppler navigation system on the aircraft and also precise data were obtained by making the flights over an Askania Camera tracking range which operates at Edwards Base. Flights were made as high as possible to give smooth flight conditions. The gravity "profile" obtained was the result of the effect of gravity along a wide (about 40 mile) strip of the Earth's surface. In the counter readings the motion of the beam was solved by an analogue computer to give the integral of the required beam correction. The results were quite satisfactory; an accuracy of better than

10 mgal was obtained which meets requirements for geodetic application where mean gravity values for $1° \times 1°$ squares are required. The average gravity values over the test area, measured at 20 000 and 30 000 ft are consistent with the average gravity over the same area on the ground, allowing for a simple free air correction.

In conclusion *Dr Thompson* said that actual application of airborne gravity measurements will perhaps be possible in the not too distant future.

## Gravity data from satellite observations

One of the outstanding developments since the last meeting of the International Gravity Commission has been the launching of artificial satellites followed by the analysis of their orbits to give new and very precise information about the Earth's external gravity field. As this was the first time that a report on this subject had been presented to the Commission, *Dr A. H. Cook* who made the report began with a section on the principles of the expansion of the Earth's external gravity field in spherical harmonics and on the theory of the field of an ellipsoid of revolution, as well as summarizing the principal perturbations to which the orbits of artificial satellites are subject due to the fact that the Earth differs, though slightly, from a uniform sphere. The mathematical methods for the study of these perturbations were outlined and the principal results so obtained were tabulated. Although all investigators obtain the same results to first order, there are discrepancies when the theories are carried to second order and these require attention.

The results so far obtained from observations of Sputnik 2, Vanguard I and Explorer 4, were summarized.

In presenting his report, *Dr Cook* explained the principal secular and long period perturbations that occur in the orbits and emphasized that the outstanding result of the work to date was that low order harmonics in the gravitational potential had been found with uncertainties at least twenty or thirty times less than those of estimates obtained from surface gravity data. On the other hand variations of gravity with longitude could not as yet be found from satellite motions.

Opening the discussion the Chairman suggested that members should concentrate on the relation of results to geodesy and geophysics. *Dr W. Kaula* then reported some results obtained by *O'Keefe*, *Eckels* and *Squires* from an analysis of the orbit of Vanguard I. Five quantities had been found from 2 000 observations extending over twelve months—namely, the secular regression of the node and motion of perigee, and the long period variations of eccentricity, longitude of perigee and longitude of node. The coefficients $J_2$, $J_3$, $J_4$ and $J_5$ had been determined; $J_2$ and $J_4$ so found from one satellite agreed extremely well with estimates obtained by combining data from Sputnik 2, Vanguard I and Explorer 4.

In a publication by *E. Buchar* presented during the meeting, the author made a first study of the movement of the orbital node of Sputnik 2 (1957$\beta$) during four months and deduced the oblateness of the Earth.

*Dr Heiskanen* mentioned that harmonics in the Earth's surface gravity field up to the eighth order had been found at Columbus and those up to the sixteenth were now being computed.

*Dr Cook* opened a discussion on the relation of data obtained from satellites to those found for surface gravity. One method, represented by papers published by *Cook*, *Arnold* and *Kaula*, was to combine the two sets of data by a

least squares method and *Dr Cook* asked *Dr Kaula* to describe his work, which was by far the most elaborate of these studies. It depends on the application of the theory of Markovian ensembles to coupled values of free air anomaly and height, and in making a spherical harmonic analysis of surface gravity data, *Dr Kaula* has included condition equations derived from satellite data. He showed copies of the geoid he had obtained from this analysis.

*Dr Cook* said that in his view although this approach was legitimate, a different one should not be ignored. Satellite results were still subject to a number of uncertainties and reasons for possible serious errors in estimates of harmonics from surface gravity were well known. In these circumstances he thought that satellite observations and surface gravity data should be reduced separately and then compared to see if discrepancies shed any light on the errors of either method.

The delegates expressed the hope that authorities launching further artificial satellites should if possible choose orbits that would yield new information about the external field of the Newtonian attraction of the Earth.

We have reported only some of the problems which were discussed by the Commission. However one should mention also the Report of *Dr de Graaff Hunter* on the computation of anomalies and the Model Earth Reduction, the catalogue of the first order stations published by the International Gravity Bureau to make easier the precise re-occupation of the international fundamental stations, and finally the National Reports and various publications presented during the Symposium. All of these will be used as a basis for new work and new discussions at the General Assembly in Helsinki in 1960.

# Geophysical Discussion of the Royal Astronomical Society

A discussion was held in the rooms of the Royal Astronomical Society on 1959 November 27 at 16h 15m on **Continental Drift,** under the Chairmanship of *Professor L. Hawkes, F.R.S.*

*The Chairman* in opening the discussion said that the concept of continental drift had been under discussion for nearly 50 years. It was of value to review the subject from time to time in the light of our continually increasing knowledge of the crustal rock strata. That portions of the Earth's crust could move tangentially over the Earth's interior had long been known to geologists from the evidence provided by large-scale tear faulting. Such movement was going on at the present day at rates which had been measured and it was to be expected that it had been in continuous operation throughout geological time—at different times in different areas. The problem of determining the direction of these past movements and the size of the moving crustal rafts was a complex and difficult one, largely due to an insufficiency of evidence, and it would be long before even a partial solution would emerge. *The Chairman* then called the speakers in turn.

The first speaker was *Dr T. F. Gaskell*, of the British Petroleum Company, who spoke in place of Sir Edward Bullard. Dr Gaskell pointed out that the seismic measurements made in the oceans since the war demonstrated that the geological structure beneath the oceans was different from that under the continents, and that the ocean basins never had been continents. When the seismic results were looked at in closer detail there appeared to be small differences, for example between the Atlantic and the Pacific oceans. In the Pacific the seismic results, supported by other measurements such as soundings and drilling, demonstrated that isolated volcanic islands followed a pattern of growth and subsequent sinking and sometimes the islands finished up as flat-topped sea-mounts or atolls. The fact that this movement of individual masses in the Earth's crust could take place suggested that the Pacific crust at least was not strong enough to resist movement. In the Atlantic on the other hand, the coral at the Bermudas was thin and there was no evidence of the sinking regime. Could it be that the crust of the Atlantic was for some reason stronger than that of the Pacific? It was possible that during some form of continental drifting the crust beneath the Atlantic was altered sufficiently to give it increased rigidity. Possibly a mechanism for horizontal movements of continents was connected with heat coming from inside the Earth. Most of the calculations of the heat flow and heat production in the mantle suggested that conditions very near melting occurred at the top of the mantle. It was possible that the blanketing effect of the crust caused a periodical overheating of the mantle with a consequent increase in fluidity. This would allow for movement of parts of the crust relative to each other and would be accompanied by enormous volcanic activity. This sort of activity was what was observed at the geological periods of mountain building. Although no definite mechanism for continental drift had

been worked out in detail, it was clear that the results from rock magnetism studies were requiring a new appraisal of the problem. The speaker's advice was for an impartial view to be adopted wherein all good evidence might be accepted and attempts made to weld it into a sensible picture. He stressed that facts relating to one part of the world should not be translated into dogmatic assertions and generalizations. The case of the Darwin–Murray controversy on coral atolls should be a lesson to all.

The second speaker was *Professor S. K. Runcorn*, of King's College, Newcastle, who concerned himself directly with the rock magnetism research. He said that if one disregarded the independent problem of reversed magnetizations the results of studies of the natural remanent magnetism of crustal rocks had to be viewed in the light of three interpretations. First the remanent magnetism vectors showed a scatter of directions apart from that attributable to experimental error. This could be explained as a reflection of magnetizations acquired at different parts of the geomagnetic secular variation pattern, and this interpretation was supported by the observation that in recently formed rock units, whose time of formation exceeded the secular variation period, the vector mean of such scattered results coincided well with the direction of the geomagnetic dipole field. The second interpretation was that of polar wandering. Meaned over periods of time which were long as compared with the secular variation, and then compared between geologic periods, the magnetic vectors suggested that the magnetic pole had undergone large displacements in the course of the Earth's history. Agreement of these results within continents was good and, when used to extrapolate pole positions, gave rise to simple tracks of polar movement. The agreement of Triassic results throughout North America was cited as a particular example of the continental agreement. The comparison of results from different continents, however, did not fit into one such simple pattern, and this led to a third viewpoint: that of interpreting the results as evidence of movement between continents. This evidence of continental drift had some broad agreements with Wegner's proposals, and the palaeomagnetic results for the ages of the glaciations observed in the geology of the southern hemisphere and of the widening of the Atlantic basin were cited as particular instances of this agreement. Several lantern slides were shown to illustrate each interpretation.

The next speaker, *Professor O. T. Jones, F.R.S.*, of Cambridge University, discussed the historical development of the concept of glaciations in the southern hemisphere and went on to plead for a modern study of some aspects of this problem. He said the concept of earlier glaciations than those occurring in the Pleistocene period developed in this country from a paper prepared by Ramsey in 1854 concerning the striated and polished pebbles of Permian deposits (now stratigraphically assigned to the upper Carboniferous) of the English Midlands. In 1855 the brothers Blandford were sent to India as geologists and in studies of the Talchir coalfield they proposed a glacial origin for a conglomerate which lay at latitude 23°N. A few years later, 1868, Sutherland published a paper on the Natal boulder beds with the conclusion that these were glacial deposits. This particularly came as a way of interpreting the association of fine and coarse material in these deposits. Sutherland also needed the picture of a heavy body moving across the rock surface with pressure. Oldham in 1874 published a restudy of the Midland Permian breccias of Ramsey's work and interpreted them not as glacial deposits but as fanglomerates similar to those forming today in central Asia. Oldham did not dismiss the striations of the pebbles, however, as evidence for glacial action

on the material some time before it was deposited. The speaker said all the features which were stressed as evidence of glacial action could be formed in three alternative ways due to gravitational slumping, mud flows, or the development of fanglomerates in arid regions. The glacial origin of the deposits in the southern hemisphere seemed strange in a time sequence which led on to coal formations and then to red sandstones. The coal contained a Glossopterus flora which was in itself no indication of climate but which was associated with plants which in the opinion of Dr J. Walton (private communication) were suggestive of a warm climate. The red beds certainly suggested tropical aridity. The speaker said the overall impression led him to favour tropical mudflows as the origin of the boulder beds of the southern hemisphere. However modern methods of research if diligently applied should settle the question in a more satisfactory way than was possible from the existing literature. Professor Jones asked for more detailed geologic mapping of the deposits, say at a scale of six inches to the mile, which should show the thickness and continuity of both the deposits and the surface on which they rested. Petrological examination of the sediment might reveal if it had formed in tropical or polar conditions. Even if the deposits proved to be of glacial origin, the speaker concluded, it must be asked if the southern hemisphere glaciations occurred in succession, due to pole movement, rather than contemporaneously; or if a world-wide decline in temperature could account for them, again with the possibility of a pole displaced from its present position.

The fourth and final speaker was to have been *Mr N. L. Falcon* of the British Petroleum Company. As this gentleman was unable to be present his address was read for him by Mr A. N. Thomas. It began with a discussion of large horizontal movements of sialic materials along tear faults within the continental masses or near their borders. The geological evidence for this rested in studies of such features as the San Andreas fault of California where the separation along the fault appeared to increase in magnitude on the two sides of the fault as successively older formations were examined, culminating in a figure of about 350 miles for the basement rocks. The Great Glen fault of Scotland, the Ocoa, Bocono′ and El Pilar faults of South America, the Bartlett fault of the Carribean, and the Alpine fault of New Zealand were among the other tear faults mentioned. With such large horizontal displacements becoming recognized more frequently by geologists, it was possible to envisage similar lines of great disturbance bounding the continental blocks beneath the oceans, helping in continental displacements but out of reach of normal geological examination. Evidence of continental similarities both in structure and stratigraphy between Africa and South America, combined with the thin oceanic crust indicated by geophysics, favoured continental drift. W. B. Harland was said to have cautiously concluded that drift best explained the major facts of regional geology in the North Greenland–Spitzbergen area.

Mr Falcon had asked other workers to summarize palaeontological evidence and studies of Gondwana sedimentation; these too were read by Thomas. *Dr F. E. Eames* summarized the palaeontological evidence as regarded three groups: terrestrial vertebrates, acquatic vertebrates other than fish, and benthonic marine faunas. Considering these on the two sides of the Atlantic he concluded that while general similarities existed at the level of genera correspondence at the specific level was rare. Falcon commented that palaeontological enthusiasm for drift seemed to have cooled. In his contribution, *Dr J. Spence* discussed the sediments of the Gondwana System as evidence for an attachment between South America, Africa, and India during that time period. He described in some detail

the Karoo sediments of South Africa and followed this by describing the Gondwana beds of South America and India. All showed thick and extensive freshwater sediments not separated from the present oceanic coast by any intervening mountains or ridges. He said these had been parts of one large freshwater basin rimmed by a mountain belt which separated the intra-continental basin from the ocean. Similarities between this basin, with the asymmetry of its structure and with the extent of its deposits, and others known in the geologic column were pointed out. This contributor strongly felt that the similarities of the history of these three continents proved the hypothesis of continental drift.

In the discussion which followed several additional points were raised. *Mr W. B. Harland* pointed out that many tillites had been examined according to the criteria put forward by Professor Jones and were acceptable. The speaker was reasonably certain that at least the late Pre-Cambrian horizons of Greenland, Spitzbergen, and Norway which he had personally examined contained good aqueous tillites. Preliminary palaeomagnetic results showed that the tillites, in conformity with the rocks of apparently higher temperature facies above and below them, suggested low geomagnetic latitudes. This was in conformity with earlier palaeomagnetic studies in Britain and North America. He was forced to conclude that whatever the magnitude of later ice ages the late Pre-Cambrian one was both intense and prolonged, and had an important effect on the nature and distribution of life. *Professor J. D. Bernal, F.R.S.*, spoke of some results reported at the International Oceanographic Congress held in New York in 1959 August and September which he felt provided additional evidence in support of the continental drift hypothesis. These results stressed features of the mid-oceanic ridges and the rift valleys which appeared to lie along their length, the magnetic and thermal anomalies of these rifts, and the absence of the Mohorovičić discontinuity in seismic studies of the ocean floor near these ridges. Bernal associated the ridges with the dividing lines of the Gondwana continent and suggested that there upwelling occurred which pushed apart the earlier crust as new crust formed. This would lead to a picture of a crust expanding in some places and contracting, under compression, in others. He felt this could explain the three major sets of deep crustal movement which he listed as: the convex arc patterns with their associated deep focus earthquakes which lay along apparent 45° thrust planes, the rift valley systems, and deep transcurrent faults. *Dr M. W. Strong* felt that the discussion lacked emphasis on the forces initiating and sustaining continental drift. In this connection he suggested some of the attractions of a modified Darwin–Fisher hypothesis for an origin of the moon from the site of the present Pacific Ocean. *Dr A. E. M. Nairn* said that two impressive anomalies existed among climatic indicators for the past: evaporite deposits in high latitudes and glacial deposits in low latitudes. Where palaeomagnetic evidence existed the latitude it predicted was in agreement with the geological indications. He described features of the Dwyka tillite of South Africa which he regarded as of glacial origin.

*The Chairman* asked if the departure of the magnetic axis from the rotational axis of the Earth might have been greater in the past. *Mr B. R. Leaton* mentioned the observed rate of westerly drift of the geomagnetic field and asked how the periodicity this implied related to the time scale of the slower pole movements. *Professor Runcorn* replied that there was evidence for relative movement between the core and mantle. In time this would average out to an axial field any non-axial

components. This time scale of hundreds or thousands of years was much shorter than that of pole movement or drift. Calling attention to the limited extent of transcurrent faults which showed slip movements rather than displacements perpendicular to the fractures, *Mr B. D. Loncarevic* asked if the faults needed for movement on a continental scale were masked, unrecognized, or non-existent. *Dr A. H. Cook* added to this that the displacements suggested by Dr Hodgson's studies of the first motion of seismographs were those of the continents rotating about the Pacific. This was not consistent with movement towards the centre of the Pacific, and the motions observed on the major transcurrent faults were not those required to bring about continental drift. In reply *Dr Gaskell* said that to look to the present fault patterns might not help much as it was possible no drift was occurring at present. Perhaps the problem of the Earth's heat regime held some answers with drift being a reflection of a cooling crust as heat was added from within the Earth. Certainly more knowledge was needed about the heat balance in the deep crust and perhaps the proposed Mohole would provide this. *Professor Runcorn* asked if the transcurrent faults were active on a global scale would not they produce continental drift.

*Professor Bernal* reported the maps made by Mason from magnetic traverses in the eastern Pacific which showed displacements indicative of extensive transcurrent faulting in the ocean floor. He asked if any oceanographers present would comment on the evidence for the mid-oceanic rifts. *Dr G. Deacon, F.R.S.*, said Bernal had already summarized well the material presented at the Oceanographic Conference. He asked *Dr A. S. Laughton* to comment on the mid-Atlantic rift. In answer to Professor Bernal, Dr Laughton reported that thirty crossings had been made by ships of the Lamont Geological Observatory, of which about half showed a single pronounced median rift, and another third showed several large rifts. A section of the valley at about 47°N was surveyed by Cambridge scientists and was shown to be at least 100 miles long and closed at one end. North of 53°N, German workers under Dietrich had found no extension of the valley. The rift was said to be characterized by a magnetic anomaly as well as by its topographic expression. If all the oceanic crust were under tension and continental drift were used to explain the rifts and tensional features indicated in all the oceans, then an overall expansion of the Earth was necessary. This was what had pressed Heezen to suggest his "blown up football theory" which had received much adverse criticism. A comment on Arrhenius's delineation of the equator by a bulge in the equatorial sediments produced by biological causes was inserted by Bernal who said this indicated no shift of the equator from its present position back into Tertiary times. *Mr A. N. Thomas* asked if the location of the deep focus earthquakes was compatible with a drift between the sial and sima. Bernal replied that these seemed to indicate planes of thrusting towards the Pacific but it seemed likely that both sial and sima were wedged out from the rifts of upwelling material. *Professor Jones* asked if the heat flow on the ocean ridges was higher than the average to conform with the picture of a convection cell. *Professor Runcorn* suggested that a study of Vening Meinesz's convection theory might be helpful and that this might be aperiodic. A suggestion of how the width of oceans in past epochs might be deduced from a study of the beach shingle formed at that time was put forward by *Professor Bernal*. A war-time study he had made suggested that the minimum size of shingle formed from some uniform hard rock like granite was a measure of the waveheight of the surf which formed it, and consequently of the fetch or width of the ocean which lay in front of it.

*The Chairman* ended the discussion with two comments. He said that while tear faulting led to a rearrangement and rotation of the crust it did not produce the separation which had been envisaged several times in the discussion. He added that if pole movement had to be entertained as well as the possibility of continental drift, the possibility of unique reconstructions of the crust of remote periods became almost impossible. *The Chairman* thanked the speakers and other contributors to the discussion and adjourned the meeting at 18h 15m.

# The First International Space Science Symposium Nice, 1960 January

The International Council of Scientific Unions has set up a Committee on Space Research (COSPAR), with members who represent countries and scientific unions, to develop international cooperation in the use of rockets and satellites for scientific observations. So rapidly have the scientific uses of these vehicles developed in the last few years that COSPAR has been able to hold the first international symposium on the results that have been obtained. It took place at Nice from 1960 January 11 to 16, under the chairmanship of Professor van der Hulst, the president of COSPAR, and was preceded and accompanied by meetings of COSPAR and its working groups. The symposium itself was attended by about 150 participants, the largest number being from the U.S.A. (65), and the next largest from the United Kingdom (36) and France (23). The Russian group was distinguished, including Professor Vernov and Professor Blaganravov, but not numerous, and there were small numbers of participants from many other countries such as Belgium, Canada, Australia, Japan and Italy, to mention just a few. Most papers naturally came from countries from which rockets or satellites had been launched or observed but there were valuable contributions from others as well. Nice, alas, did not favour us with her warmest weather, except for members of COSPAR who arrived in advance of the symposium, and was rather wet and cold: there was even snow on the promenade on two days. But the French reception committee and the town council of Nice gave us a warm welcome.

On the Sunday preceding the symposium some seventy of us visited the Observatoire d'Haute Provence, near Forcalquier. This observatory, part of the Centre Nationale de Recherche Scientifique, contains the largest telescope in Europe (84 in.) and has a comprehensive range of spectroscopic equipment, including photoelectric apparatus just being developed. Another telescope has an objective prism and another is provided with a photomultiplier of M. Lallemande's advanced type for six-colour photometry. We returned through Fréjus and saw a little of the destruction there.

An exhibition of instruments and results had been organized in the symposium building, with contributions from the U.S.A., U.S.S.R., Canada and Japan, but by far the largest and most interesting part, outweighing all the others together, was that from Great Britain in which rocket instruments and results, work on tracking, and the progress of the Manchester–Pic du Midi Moon survey were all on show.

The symposium was organized into sessions dealing with the Earth's atmosphere, cosmic radiation and interplanetary gas, the Moon and the planets, tracking and telemetry, the ionosphere, solar radiation and meteorites. Simultaneous sessions were sometimes unavoidable and on that account very little is said about the Moon and the planets nor about meteorites in this report.

The first paper was one of the most striking of the symposium. Dr H. Wexler showed a film of pictures of the Earth taken from an Atlas missile launched southeastwards from Cape Canaveral and reaching 1 000 km; a front to the north,

organization of clouds over hundreds of miles in the tropics, and the tropical convergence zone were clearly revealed in the cloud patterns and for the first time large-scale organization of clouds has been shown. The remaining papers in the section on the Earth's atmosphere were principally about measurements from rockets and satellites of pressure, temperature, composition, and winds. Among the more interesting techniques is the observation of a cloud of sodium ejected from a rocket; as is well known the sodium is excited and emits the yellow doublet and observations of the drift of the luminous cloud reveal winds, while by photographing the cloud through a Fabry–Perot interferometer the temperature can be estimated from the widths of the interference fringes. It was clear from this session that there is a real discrepancy between air densities found from rocket soundings and those found from the drag on satellite orbits, for the latter show little sign of the variations with season and latitude that are so prominent in the rocket results. The reason for this discrepancy has still to be found.

The largest group of papers and the greatest body of new information were undoubtedly in the section on cosmic radiation and interplanetary gas, for this included the discovery and study of the van Allen radiation belts, the twin regions or energetic charged particles trapped by the Earth's magnetic field. It is sometimes objected that the money spent on "space research" is out of all proportion to the results achieved but the van Allen belts were quite unsuspected prior to satellite observations and one can scarcely doubt that there are many other real unexpected discoveries yet to be made, especially in astronomy. It is also worth mentioning that van Allen's discovery was made with very simple apparatus—indeed one of the impressions of this symposium is that while it is very expensive to put a satellite into orbit, the instruments carried are in general rather simple and inexpensive; the scientific part of space research is not very costly.

Russian and American results of observations of the intensity, nature and energy spectrum of the charged particles in the radiation belts were given in a number of papers, including measurements made from the Luniks and Pioneer space probes as well as from satellites. In general, these results seem to agree though there are discrepancies in detail, but it was surprising to find what a very firm picture of the radiation belts had been built up mainly from the results of two Luniks and two Pioneers. One suspects there is a great deal more detail to be filled in, about variations of number of particles in space and time, and about the energy spectra. There seems little doubt that the present picture is oversimplified. There are large areas of the belts where no observations have been made. One also felt that all the work on the belts was at present purely descriptive. An important aspect of the study of the belts is the measurement of the Earth's magnetic field in them. Both satellites and space probes have been equipped with magnetometers for this purpose and interesting results about the form of the main field have already been obtained. Russian results have also demonstrated that any magnetic field of the Moon is probably less than $100\gamma$.

The radiation belts affect the Earth in various ways, notably by supplying energy to the top of the atmosphere, thus changing the general circulation, and by producing alterations in the magnetic field at the surface of the Earth.

Three pictures of the reverse of the Moon taken from Lunik III were on view in the exhibition and Professor Blaganravov described the technique used to obtain them. This was of most concern to the engineers but it was very interesting to hear Professor Blaganravov say that contrast had been sacrificed in order to obtain a photograph of the entire surface of the Moon.

There were a number of interesting papers on problems of tracking, especially of Moon probes and there were two groups of papers of especial interest to geodesists. One included papers on the theory of the orbits of artificial satellites and the derivation of the Earth's gravitational field from the changes in the orbits. It is now well known that the low-order spherical harmonics in the expansion of the external potential of the Earth can in principle be obtained far more accurately by this means than from, say, gravity observations on the surface of the Earth, although the details of the theory are still subject to discussion and the variation with longitude cannot at present be found. But this high precision is already stimulating new approaches to the theory of the Earth's field and to the geophysical significance of the results that are being obtained. While the van Allen belts are a quite new discovery, the new interest in the Earth's gravitational field shows how an old subject may be rejuvenated by a new means of observation. Geometrical geodesy will also benefit from rockets and satellites. Nothing was said at this symposium about satellites with flashing lights but the possibilities of tracking and geodetic measurements by means of observations of the Doppler shift of radio signals from satellites were discussed in two papers.

The ionosphere has been a prime subject of study from the very first, by means of rockets as well as satellites, and its study, like that of the Earth's gravitational field, has shown that important results can be obtained from a careful analysis of observations made from the ground, in this case of refraction and Faraday rotation of radio transmissions from satellites.

Perhaps the most impressive group of papers, alike for the results already obtained and for the promise of future discoveries, was that concerned with ultraviolet spectroscopy with rocket-borne instruments. Already a great number of emission lines in the spectrum of the Sun has been discovered, the dominant ones being Lyman-$\alpha$ and the corresponding line of singly-ionized helium; the profile of solar Lyman-$\alpha$ has been obtained and the Sun has been photographed in Lyman-$\alpha$. A spectrum of Lyman-$\alpha$ obtained just before the symposium shows intense absorption by neutral hydrogen in the Earth's outer atmosphere, the first unambiguous evidence of its presence. Resonance scattering of Lyman-$\alpha$ from hydrogen around the Sun has also been observed. At the same time it is clear that atmospheric absorption still colours spectra obtained from rockets and there is undoubtedly a rich harvest to be gathered as soon as spectroscopic observations can be made from satellites. Two technical advances are necessary for this to be done—the development of telescopes stabilized with respect to the stars and the development of spectroscopic methods using photoelectric recording and telemetric transmission of data instead of photographic methods; the first such photoelectric system, used in a rocket, was described in one of the papers in this group.

Altogether this was a most interesting symposium and at times exciting. A number of impressions stand out. The first, unfortunately, is that many of the papers were poorly delivered. Many authors tried to say too much in the time available without first explaining the background and significance of their work. Linked with this was the fact that the discussion of papers was usually brief and sketchy. It seemed significant that the best discussions were on two subjects—physical geodesy and solar spectroscopy—where there is an established body of theory and observation to provide a context and a critique for the new material. In so many other fields one was presented with a great deal of descriptive material, presented in detail, it seemed, because the theory that would allow it to be summarized had not yet been developed.

A number of participants felt that it was useful to have this symposium in which methods and results in diverse fields were brought together and that perhaps a further such symposium might be justified. It is certainly useful, for instance, to learn about other people's techniques when the instruments are still so novel, but the disadvantages of this symposium were also clear: with students of so many different subjects gathered together, scarcely any topic was discussed at all deeply. It seems that, more and more, astronomical results should be discussed in the International Astronomical Union, geophysical results in the International Union of Geodesy and Geophysics, and so on.

1960 *January* 27.

# Book Review

## Fonctions Sphériques de Legendre et Fonctions Sphéroïdales

Louis Robin

(Tome III 1959, 289 pp. + viii, 5,500 FF. Gauthier-Villars, Paris)

The first two volumes of this work were reviewed in *Geophys. J.*, **2**, 259, 1959. This third volume completes the book.

In Chapter VII formulae for the addition of Legendre polynomials are extended to the general functions $P_n(\mu)$ and $Q_n(\mu)$, where $n$ and $\mu$ are any numbers, real or complex. Additional formulae are also given for $P_n^m$. Chapter VIII deals with zeros of $P_n^m(\mu)$ and $Q_n^m(\mu)$, regarded as functions of $\mu$, for real $m$, $n$, and with allied problems. In Chapter IX the author makes elegant applications of the associated Legendre functions to potential problems for surfaces of revolution, considering spheroids, toroids, and the infinite half-cone, with applications of the Kelvin inversion theorem. Chapter X treats spheroidal wave functions, Gegenbauer polynomials and Gegenbauer functions. Appendix I deals with spherical Bessel functions and Appendix II lists numerical tables of Legendre functions existing at the end of 1958.

Throughout the work one finds mention of previously unpublished material or corrections of published errors. The high standards of rigorous statement and manipulative skill which mark Volumes I and II are maintained in Volume III. The treatment is full and detailed, yet the main characteristics of the functions under consideration are never lost sight of. Whether used as an exposition or as a work of reference, these three volumes should give great satisfaction to the reader.

E.R.L.

## РЕЗЮМЕ ДОКЛАДОВ, ОПУБЛИКОВАННЫХ В No. 4 ВЫПУСКЕ ПРЕДЫДУЩЕГО ТОМА В ПЕРЕВОДЕ НА РУССКИЙ ЯЗЫК

### РАЗМАГНИЧИВАНИЕ НЕСТАБИЛЬНЫХ ТРИАСОВЫХ КЕЙПЕРОВСКИХ МЕРГЕЛЕЙ ИЗ ЮГО-ЗАПАДНОЙ АНГЛИИ ПРИ ПОМОЩИ ПЕРЕМЕННОГО ТОКА

*К. М. Крир*

В 1953 г. было изготовлено 329 маленьких дисков из 35 образцов горных пород, собранных из 60-метровой толщи кейперовских мергелей (триаса) в Сидмус (Девоншир). Результаты анализа направлений естественного остаточного намагничения образцов, описанные Криром (1955, 1957а), показали существование трех раздельных составляющих намагничения, так наз. «первичное», «вторичное» и «временное». Повторные измерения 65 экземпляров этих дисков, относящихся к 26 из 35 рассматриваемых образцов пород, выполненные после 3-летнего хранения образцов в свободных полевых условиях, показали, что «временная» составляющая намагничения продолжала ослабевать. В дальнейших опытах эти же 65 дискообразных образцика были размагничены в переменном магн. поле с амплитудой в 1200 эрстеда. Направления намагничения, вытянутые первоначально вдоль большего круга между направлениями первичного и вторичного намагничения, по мере увеличения размагничивающего поля постепенно собираются в две группы с противоположным намагничением. Оказалось, что во многих случаях вторичное намагничение почти полностью уни чтожается, а в остальных случаях значительно ослабевает по сравнению с первичным. Описан метод размагничивания переменным током.

### АНОМАЛИЯ ГЕОМАГНИТНЫХ ВАРИАЦИЙ В ЯПОНИИ

*Т. Рикитакэ*

По данным наблюдений магнитных обсерваторий Японии, хорошо распределенных по территории страны, описаны аномальные свойства различного рода медленных и быстрых геомагнитных вариаций. Так как эти аномалии вызываются индуцированным магнитным полем внутреннего происхождения, то рассмотрены ответственные за аномалии физические условия земных глубин. В настоящее время лучшим путем объяснения аномалий представляется допущение, что проводящая цепь, поднимаясь из глубины, образует под Японией петлю примерно в 1000 км длиной и 200 км шириной. Предполагается также, что здесь имеет место клинообразная интрузия верхнего слоя мантии. Глубина основания проникшей части оценивается в 700 км. Эта сложная структура имеет большое значение для геофизической интерпретации связи между глубокофокусными землетрясениями, орогенической активностью, распространением вулканических зон и т.д.

### РАСПРЕДЕЛЕНИЕ КАСАТЕЛЬНЫХ НАПРЯЖЕНИЙ В ПРИЛИВНОМ ТЕЧЕНИИ

*К. Ф. Бауден, Л. А. Фэрбэрн, и П. Хьюз*

Наблюдалось приливное течение мористее бухты Ред-Уорф, Англси, Сев. Уэльс. Напряжение трения у дна, $F_b$, вычисленное по профилю скорости в пределах первых 2 м над дном находится в квадратической зависимости от скорости на глубине 1 м от дна, $F_b = k\rho U_1^2$, где $k$ имеет значение $3{,}5.10^{-3}$. Отмеченные значения шероховатостей имели длину 0,16 см. Течения измерялись с интервалом в 1/2 часа, что дает возможность определить члены ускорения в уравнении движения. По напряжению на дно и членам ускорения вычислено полное касательное напряжение как функция глубины и как функция времени в течение приливного периода. Во время наступления максимума

течения значения касательного напряжения увеличиваются приблизительно линейно от поверхности ко дну, как в случае установившегося потока в канале. В остальное время члены ускорения вызывают напряжения значительно отличающиеся от линейных колебаний. Вычисленные значения вертикальной турбулентной вязкости, $N_z$, показывают некоторое увеличение от поверхности и дна к середине глубины. $N_z$ изменяется в течение приливного периода, достигая максимума во время максимума течений и будучи выше при приливе, чем при отливе. Численные значения $N_z$ были порядка 270 см²/сек при приливе и 130 см²/сек при отливе, соответствуя скоростям течения в середине потока 35 см/сек и 39 см/сек. Средняя глубина воды 22 м. Наблюдаемые распределения скорости и касательных напряжений сопоставлены с полученными теоретическим путем в котором турбулентная вязкость взята постоянной над слоем трения вблизи дна.

## ПАЛЕОМАГНЕТИЗМ ГРЕЙТ ВИН СИЛЛА

*К. М. Крир, Э. Эрвинг и А. Э. М. Нэрн*

Образцы интрузивных пород Грейт вин Силла, относящихся к периодам от Верхнего каменноугольного до нижнего пермского, собраны в 36 местах в Нортумберленде и Дургаме. Они оказались магнитно стабильными, со средним направлением намагничения 187.8°—4.9°; соответствующее ему положение полюса 37.3° С, 168°.9 В. Распределение средних направлений намагничения для этих мест не круговое, а овальное, причем изменения в направлении палеомагнитных долгот вдвое больше, чем в направлении палеомагнитных широт. Однако, положения полюсов, вычисленные из средних направлений намагничения, имеют круговое распределение; в связи с этим предложен метод вычисления ошибок, связанных с определением местоположений палеомагнитных полюсов.

## ГРАВИТАЦИОННЫЕ АНОМАЛИИ И СТРУКТУРА ЗЕМНОЙ КОРЫ В ЮЖНОЙ АФРИКЕ

*А. Л. Хейльс и Д. Я. Гоф*

В этой статье дан расчет поля изостатических аномалий Южной Африки. Оценка мощности земной коры производится путем использования критерия коэффициента регрессии, предложенного Хейскаценом, критерия минимальной суммы квадратов, предложенного Бауй и нового критерия, основанного на минимальной сумме квадратов отклонений от линии регрессии. Найдено, что критерий коэффициента регрессии недостаточен, так как он чувствителен к систематическому эффекту, проявляющемуся в Южной Африке около вертикальных выходов пород. Другие критерии дали результаты, согласующиеся со строением земной коры, полученным с помощью сейсмических исследований, а именно мощность земной коры определена в 30-40 км.

## НАБЛЮДЕНИЯ С ПОМОЩЬЮ КЕМБРИДЖСКОГО МАЯТНИКОВОГО ПРИБОРА В СЕВЕРНОЙ, ЦЕНТРАЛЬНОЙ И ЮЖНОЙ АМЕРИКЕ В 1958 Г.

*Дэ. Е. Джэксон*

Наблюдения с помощью Кембриджского маятникового прибора были проведены в августе-декабре 1958 г. в Теддингтоне, Мадисоне, Мехико-сити, Панаме, Каракасе, Квито, Лиме, Ла-Пасе, Сантьяго (Чили), Буэнос-Айресе и Рио-де-Жанейро. При обратном ходе стоянки приборов были расположены так, что между двумя соседними станциями было произведено восемь измерений. Полученные величины рассмотрены в статье.

## СКОРОСТИ ВОЛН ПОД ГРАНИЦЕЙ МОХОРОВИЧИЧА

*Б. Гутенберг*

Скорости продольных волн непосредственно под границей Мохоровичича, найденные различными авторами для континентальных районов, уменьшаются от значений 8,2 км/сек. при глубине залегания границы Мохоровичича, равной 30 км, до 8,0 км/сек.,

если ее глубина 50 км. Градиент убывания скорости, уверенно определенный статистически, равен 0,011 ± 0,0016 км/сек. на километр и превышает максимально допустимый для объяснения существования канала с малой скоростью градиент 0,0013 км/сек. на километр. По вычисленному уменьшению скорости поперечных волн от 4,64 км/сек., при глубине залегания границы Мохоровичича 30 км, до 4,56 км/сек. для глубины 50 км (градиент 0,004 ± 0,0056 км/сек. на километр) можно ожидать средних условий под континентом, в случае, если это уменьшение в скорости рассматривается в связи с соответствующим ее уменьшением для продольных волн. Для глубин в 60 км обе вычисленные кривые «скорость — глубина» примерно совпадают с найденными автором для кажущихся скоростей в точке излома годографов для землетрясений, возникающих на различных глубинах. Новые результаты соответствуют гипотезе о том, что канал с низкой скоростью начинается от границы Мохоровичича.

## ПАЛЕОМАГНИТНОЕ ИССЛЕДОВАНИЕ НЕКОТОРЫХ НИЖНЕЮРСКИХ ГОРНЫХ ПОРОД В СЕВ. ЗАП. ЕВРОПЕ

*Р. В. Гердмер*

Проведено исследование направлений естественного остаточного намагничения пород песков верхнего лиаса в Англии (раздел 1) и приведены доказательства магнитной стабильности некоторых из этих пород. Результаты показывают, что направление магнитного поля во время верхней юры существенно отличалось от современного поля и имело восточное склонение и наклонение 64° ÷ 70°. Исследование было распространено (раздел 2) на некоторые нижнеюрские вулканические породы из южной Франции; они дали результаты в общем схожие с результатами, полученными для Англии. Сделано заключение (раздел 3), что геомагнитный полюс в период нижней юры располагался в районе Каспийского бассейна.

## О СВЯЗИ МЕЖДУ СКОРОСТЬЮ ВЕТРА И АМПЛИТУДОЙ МИКРОСЕЙСМ НА БЕРЕГУ АНТАРКТИКИ

*Дж. Мак-Дауалл*

На базе Королевского общества в заливе Халли в период летнего сезона, от декабря до февраля, наблюдались значительные микросейсмы, возраставшие во время ветра. Даны примеры наблюдений за микросейсмами и ветром в период с 30 декабря 1957 г. по 6 марта 1958 г., иллюстрирующие природу связи между двумя ними и показывающие быстрое убывание микросейсмической активности около конца февраля.

# Geophysical Journal

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# Contents


| | PAGE |
|---|---|
| D. S. Parasnis, The compaction of sediments and its bearing on some geophysical problems | 1 |
| P. N. S. O'Brien, Seismic energy from explosions | 29 |
| Mohammed Aftab Khan, The remanent magnetization of the basic Tertiary igneous rocks of Skye, Inverness-shire | 45 |
| M. H. P. Bott, The use of rapid digital computing methods for direct gravity interpretation of sedimentary basins | 63 |
| Leon Knopoff & Gordon J. F. Macdonald, An equation of state for the core of the Earth | 68 |
| J. E. Jackson, Pendulum observations at Teddington, Singapore, Darwin and Melbourne in 1959 | 78 |
| *Notes on Progress in Geophysics* | |
| Brig. G. Bomford, The figure of the Earth—its departure from an exact spheroid | 83 |
| E. Irving, Palaeomagnetic pole positions, Part I. Pole numbers 1/1 to 1/144 | 96 |
| A. Decae, On some movements of the ground in Geneva | 112 |
| *Reports of Meetings* | |
| Meeting of the International Gravity Commission, Paris, 1959 September 15–19 | 121 |
| Geophysical Discussion of the Royal Astronomical Society, 1959 November 27 | 127 |
| First International Space Science Symposium, Nice, 1960 January 11–16 | 133 |
| *Book Review* | 137 |
| Russian translations of summaries of papers in *Geophys. J.*, 2, No. 4 | 138 |
| РЕЗЮМЕ ДОКЛАДОВ, ОПУБЛИКОВАННЫХ В No. 4 ВЫПУСКЕ ПРЕДЫДУЩЕГО ТОМА В ПЕРЕВОДЕ НА РУССКИЙ ЯЗЫК | 138 |


*Printed in Great Britain by J. W. Arrowsmith Ltd., Bristol*